COASTAL AND SUBMARINE MORPHOLOGY

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ANDRÉ GUILCHER

Professor of Oceanography at the Sorbonne

translated by

B. W. Sparks and the Rev. R. H. W. Kneese

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INTRODUCTION

The two parts of the earth's surface studied in this work are very unequal in size. If we define the coastal zone as the intertidal zone it covers, according to Ph. H. Kuenen, only 150,000 sq. km., but the exact area is difficult to calculate as a result of estuaries. One of the greatest widths exposed by the tide, 20 km. or more, is found in the bay of Mont-Saint-Michel.

In practice, the zone affected by coastal processes is somewhat greater than this, since we have to consider cliffs and an area below low water springs, the exact extent of which is a matter for discussion (pp. 19 2[illegible] and 69). Thus, we are dealing with the zone where subaerial, submarine, and purely coastal processes interact.

The submarine zone, if we consider its area, is of the greatest importance in geomorphology, since it covers 70·8 per cent. of the earth's surface, excluding lakes. Its actual significance is not yet proportional to its area, because direct visual study below 40–50 m. is still difficult, dangerous, and rarely undertaken. Except in shallow water, all methods of investigation are 'blind'. Although this state of affairs will probably be changed in the near future, the coastal zone is at present infinitely better known than the submarine zone. This is the reason why we shall discuss it in some detail, but only give a general outline of submarine geomorphology. Thus, the balance between the two parts of this work represents the present state of knowledge and is intended as no slight on the value of submarine research, which is as difficult as and more costly than research in deserts and high mountains. We sincerely hope that the submarine section of this book becomes rapidly out of date, and that before very long it will be possible to give a fuller, more precise, and less hypothetical, but still geographical, account of the sea-floor.

I wish to express my sincere thanks for help, information, and advice, which they freely gave during the preparation of this work, to my teachers MM. de Martonne and Cholley as well as to MM. Baulig, Berthois, Birot, Bourcart, Cailleux, Meynier, and Tricart, to the chief

hydrographic engineers MM. Gougenheim and Lacombe, to M. de Rouville, Inspector General of Lighthouses and Buoys, to Professor Fairbridge of Columbia University and to all others who have helped.

As the French edition of this book appeared in 1954, modifications and additions have been made here in the text and in bibliographies. I have been helped by the criticisms of Professors Baulig and Musset, and I have incorporated the results of some recent advances in knowledge, so that this is really a new edition and not merely a translation. Many thanks are due to Professor Steers, the originator of the English edition, and to Mr Sparks and the Rev. Kneese, who have prepared the translation.

A. G.

PART ONE

Coastal Geomorphology

Chapter I

THE FORCES IN ACTION

A. WAVES

There is no point, in coastal geomorphology, in dwelling at length on waves in deep water, as these neither affect the sea-bed nor are affected by it. A few general points about such waves will suffice. Swell, which is produced by the wind but travels beyond the area where it is formed, is not affected by the sea-bed, as long as the depth is greater than approximately half the wave-length, i.e. the distance between two successive crests. Wave-lengths of more than 200 m. are rare. On the other hand, the orbital motion, imposed upon the molecules of water by the swell, does not cause them to return to exactly the same positions, because the radius of rotation decreases with depth. Thus, the backward motion in the troughs is less than the forward motion on the crests. Swell, therefore, creates a weak current in the same direction as itself, as Gilbert realized in 1885. Finally, the size of the swell depends on the speed and duration of the wind and the fetch, or distance over which it has blown. The increase in size of the waves is rapid at first but much slower when the fetch exceeds 1,000 km. and the time is longer than 50 hours (Sverdrup and Munk).

Waves near the coast. Near the coast swell undergoes two types of deformation before breaking.

(*a*) *In direction.* It suffers the same types of deformation as light rays—refraction, reflection, and diffraction, because it is a wave phenomenon.

Refraction is a change of direction due to the effect of the sea-floor and occurs when the depth is less than half the wave-length, provided that the isobaths are not parallel to the waves. Its extent may be calculated if we know the direction and length of the initial waves, the period of the waves, i.e. the time between the passing of successive crests, and the relief of the sea-bed, provided that the slope of the bed is not greater than 1 in 10. The refraction formula is $\frac{\sin \alpha}{\sin \alpha_0} = \frac{c}{c_0}$, where α is the angle

between the wave crests and the isobaths at a given depth, c the speed at that depth, and α_0 and c_0 the values for the same in deep water. The result is that waves tend to become parallel to the isobaths, although, where they start very obliquely, they remain to some degree oblique right up to the shore. Irregular sea-floors give unexpected results, such as concentrations of crests at certain points on the coast (Fig. 1 A). A submarine valley perpendicular to the waves causes them to diverge and thus disperse their energy. A submarine ridge perpendicular to the waves causes a convergence of the waves and hence increases their attack. Isolated swells in the sea-floor cause the waves to converge in their lee.

Reflection is the reversal of a wave by some obstacle such as a jetty or a strongly sloping foreshore (Fig. 1 F). As in optics, the angle of incidence equals the angle of reflection. Reflection of swell approaching perpendicularly causes a series of standing waves to be produced by the interference of approaching and reflected waves (*clapotis*). If the waves approach obliquely, the approaching and reflected waves form a grid pattern. A vertical obstacle causes total reflection. The degree of reflection decreases with the angle of the obstacle and is negligible when the slope is very gentle.

Diffraction (Fig. 1 F) occurs when a wave passes the end of an obstacle. The waves change direction and rapidly die out. Thus, although there is shelter in the lee of the obstacle, that shelter is not complete.

(*b*) *Apart from the changes in direction*, the wave undergoes a series of other changes before it breaks. One characteristic, the period, remains constant. The other characteristics vary with the ratio H/λ, where H is the depth at any point and λ the wave-length in deep water. The following changes occur:

1. Decrease of the wave-length.
2. Decrease of the speed of the wave.
3. Increase of the steepness of the wave (ratio between the height of the wave and its length).
4. The wave system is simplified as short irregular waves which often complicate the swell in deep water, die out or are greatly reduced.
5. The orbital paths of the molecules are changed to ellipses: near the bottom the elliptical motion virtually becomes a to-and-fro movement.

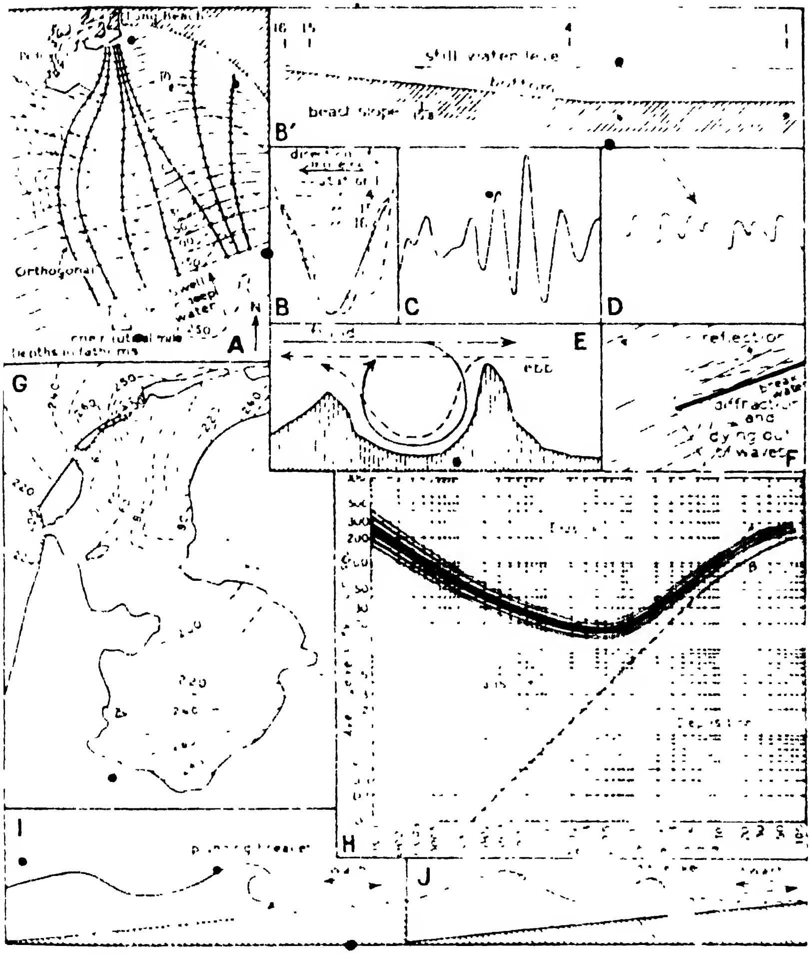

FIG. 1 WAVES AND CURRENTS

A. Refraction and concentration of swell responsible for the destruction of Long Beach breakwater (after the Beach Erosion Board, 1950). B. Wave deformation caused by shallowing of the type shown in B¹ (after Weigel, 1950). C. Variation in height of successive waves at Arguello Point, California (after Folsom, 1949). D. Zig-zag movement of material on beach caused by oblique waves. E. Tidal currents always flowing in one direction in a bay. F. Reflection and diffraction of swell by a breakwater. G. Deformation of the sea surface by a storm in the Zuyder Zee (after Lorents and Thijsse, in Kuenen, 1950). H. Hjulstrom's curves (1935, p. 298). Logarithmic scales. I. Plunging breaker. J. Spilling breaker.

6. The profile of the wave becomes increasingly asymmetrical, the front becoming progressively steeper. This is caused by the decrease in wave speed, which affects the front of the wave before the back. The back tends to overtake the front and so causes overturning.

Breaking is not directly caused, as has been stated, by friction with the bottom. It seems to result from an excess of steepness, the maximum theoretical steepness being 0·14 according to Stokes. Breaking usually occurs when the ratio between the depth of the water and the height of the wave is between 1·1 and 1·5, according to the form of the wave. All waves do not break at exactly the same point, because waves in a series are of unequal heights, as the swell consists of several series of waves superimposed on each other (Fig. 1 C). The result is that waves break not along one line but in a zone. There are two types of breaker: the plunging breaker (Fig. 1 I) and the spilling breaker (Fig. 1 J). In the former the crest plunges vertically downwards: in the second, foam appears on the crest which steadily rides over the advancing waves. Thus, in this type, the breaking occurs over a wider zone than in the plunging breaker. Plunging breakers appear to result from regular unrefracted swells of no great steepness, breaking on a smooth bottom: spilling breakers result from waves of considerable steepness and commonly occur where strong winds are blowing onshore. They are analogous with the white horses of deep water.

The wave, after breaking, is changed into a wave of translation, the swash. It flows turbulently up the shore, but is retarded by the backwash of the preceding wave, by gravity and by infiltration especially when the sand of the beach is not saturated. The backwash is far less turbulent than the uprush. The swash is thus divided into an uprush and a backwash.

Especially in storms, when spilling breakers predominate, and on beaches, such as those of the Landes of Gascony, where the waves are usually powerful, there may be two or three successive zones of spilling breakers. The wave of translation breaks again after its first uprush.

The more gradual the slope of the beach, the smaller is the height of the breakers, other things being equal. Thus, on beaches with concave profiles, the breakers are smaller at low tide than at high tide (cf. pp. 84–5). On rocky coasts, where the profile may be very different, exceptions which confirm this rule are found; for example, the highest breakers may occur at half-tide if the slope is steepest at that level, as

in the island of Sein (Guilcher, 1951). On oceanic atolls where the slopes are very steep (cf. p. 120) breakers are very high because there is practically no retardation.

The breaking wave erodes in a number of ways: it digs into the beach as it plunges down; it rolls and sorts material in the swash; takes material into suspension in breaking and the uprush; deposits material at the end of the uprush; and causes sheet erosion in the backwash. The effects on the beach will be examined in detail in Chapter III. The details of suspension are unfortunately not well known. The American *Beach Erosion Board* found 17 per mille of sand in the water in the zone of breakers and only 3 per mille some 40 or 50 m. farther out to sea (Kuenen, 1950, p. 262). King found that on English beaches the layers of sand stirred up by waves did not exceed 4 cm. except in storms. Shells are moved much more readily than sand: according to Menard and Boucot, the shells of terebratulid brachiopods are moved by currents only one-tenth as powerful as those required to move pebbles of the same size.

Waves breaking against a cliff cause erosion by variations of pressure, by impact, by hurling pebbles against the cliff and by spurting upwards. Pressure can be recorded by dynamometers (Johnson, 1919): pressures of 30 tons per square metre have been recorded, but such pressures are very local and short-lived. They act both on the rock itself and on air compressed within the rock. As the pressure is released, a suction effect is felt. Impact is capable of moving enormous masses: a block of concrete weighing 2,600 tons was moved by the waves from Wick harbour in 1877. Impact can also erode unconsolidated cliffs, but its effect on compact, non-jointed rock is more doubtful: Gilbert (1885, p. 81) considered it negligible. We will return to this subject later (pp. 62–76). Bombardment by pebbles is certainly much more effective both on the cliff and on rock exposed in the foreshore. Water spurting upwards increases the vertical range of wave action, but can be reduced by appropriate constructions which induce partial reflections of the waves. The action can extend as high as 60 m. or more: a block weighing 20 tons was lifted 4 m. by such action at Ijmuiden in Holland, while small pebbles can be flung up as high as the water itself. In storms blowing onshore, the spurting water may be beaten down on to the top of the cliff by the wind. at Belle-Ile in Brittany, sizable masses of water are hurled over a zone about 50 m. wide at a height of 40–50 m.

At a depth of one-half the wave-length, the movement in the water is $\frac{1}{23}$ of its value at the surface and gravel can still be moved by large

waves. At a depth equal to the wave-length it is only $\frac{1}{512}$. Vaughan Cornish (1898, pp. 531–2) believed that appreciable effects were caused by certain waves down to a depth of 270 m., while Johnson quoted the fact that stones, a pound in weight, were flung into lobster pots at a depth of 54 m. off Land's End. But it seems that even at depths of only a few metres, according to the careful observations made by Berthois, there is little effect on material larger than gravel (cf. p. 69).

Tsunamis, incorrectly called tidal waves, are waves resulting from submarine earthquakes. They are isolated waves and not part of a system. They travel at high velocities (of the order of several hundred kilometres an hour) and result in a series of breakers at the coast, which may last hours or even days, with a periodicity of 8–100 minutes (cf. the period of normal waves which is only a few seconds). Such waves may cause severe erosion to a height of 40 m. Similar waves are produced by masses of land or more frequently by the ends of glaciers collapsing into the sea. Such waves are of the greatest importance in the fjords of Greenland (Boyé, p. 12) where the fetch is insufficient for the wind to produce large waves.

Surf beats are waves observed sometimes on certain beaches with a period of about 2 minutes as a rule. They are superimposed on the general wave pattern and cause a displacement of the swash zone. In some ways, they resemble bores. They are probably due to the breaking of several large waves, followed by a calm.

B. CURRENTS

Wave currents. We have seen that the swell forms a current in its own direction in deep water. Breaking of waves also moves water towards the coast and has been accurately measured by floats. This water must return and it has usually been said to return along the bottom (Vaughan Cornish, Rivière, and especially Gilbert, who attributed an important role in coastal evolution to such movement). However, doubts have been cast on the existence of this undertow, and Davis (1925) gathered evidence against it. Recently Shepard and other Californian workers (1941, 1950) have shown that the water returns as rip currents, at least in certain cases. Such currents are localized in certain parts of the beach, but the localities change: they are only 15–30 m. wide, quite rapid, sometimes exceeding 2 knots, and pass through the zone of breakers. They are fed by water moving along the beach and expand and die out at their ends, while water returns up the beach between

them. They carry a much greater load than the water around them. They affect the whole depth of the water and not only the bottom. Their strength depends on the height of the breakers. Further research is necessary to understand their distribution and location. It has not been proved that an undertow does not exist, but where rip currents have been closely investigated, as at La Jolla in California, there is no undertow.

Coastal currents or *coastal drift*, unlike wave currents, run parallel to the shore and not perpendicularly to it. They result principally from oblique waves. We have seen that such waves are refracted but remain to some extent oblique. An obliquity of more than 10° is rare. With such a degree of obliquity and large breakers the coastal current can attain a velocity of several knots. Such a current may be an important agent of transport. In the swash zone material may be moved in a series of zig-zags[1] (Fig. 1 D), while movement of disturbed water also occurs in the zone of breakers.

Tidal currents have higher velocities near the shore than in the open sea and thus deserve our attention. They are especially strong in two types of locality:

(*a*) In straits where the time of the tide is very different at each end. There can, therefore, be a strong current even if the tidal range is small, as in the strait of Messina where the current reaches 5 knots even though the tidal range in the Ionian and Tyrrhenian seas is only 20–30 cm.

(*b*) In straits where the tidal current is canalized even if there is little difference in the time of the tide, provided that the tidal range is great. Such effects occur in the narrow openings of bays which widen within, such as the Rade de Brest, the Etel river and Morbihan. The highest velocities known, of the order of 16 knots (Kuenen, 1950, p. 67), are found in the Moluccas. The highest on the French coast is found in the Raz Blanchart, between Alderney and the Cotentin peninsula, where it reaches 10 or 11 knots. It appears to be caused by the projection of the Cotentin peninsula, by the differences in the time of the tide between La Hague and Saint-Malo, and by the narrowness of the strait: it does not coincide with the greatest tidal range found in the Channel, which occurs in the bay of Mont-Saint-Michel.

The fact that tidal currents change direction, four times a day in the case of semi-diurnal tides and twice a day with diurnal tides, greatly

[1] This is really a movement produced by waves rather than a direct action of currents [Translator].

decreases their effect. They can build only temporary forms, unless the forms are parallel to currents flowing in diametrically opposite directions. This is the case in straits and near the coast, but out to sea the current is usually of a rotatory nature.

Some tidal currents always flow in one direction, especially in bays, although the current may reverse in the open sea outside (Fig. 1 E). These fixed eddies, the velocity of which varies with the state of the tide, are caused by the shape of the headlands on either side of the bay: one of the headlands deflects the tidal current, while the other does not have the same effect on the current when it reverses. Very occasionally, in straits which are filled by the sea only shortly before high tide, the current is always in one direction, because the times of reversal of the current and of high tide do not coincide: it may reverse before the strait is filled so that there is always an ebb current, as at Kilaourou on the island of Sein (Guilcher, 1951), or it may not reverse until the strait is abandoned by the falling tide so that there is always a flood current.

Tidal currents in estuaries are important because the tidal range is often increased there. On the St Lawrence it increases from 2 m. at the eastern end of the island of Anticosti to about 5 m. between the river Saguenay and Quebec. Although such currents reverse direction, the ebb always carries away more water than the flood brings up as a result of the water brought in by rivers. In fact, a floating object undergoes a series of shorter movements upstream and longer movements downstream, the net result being that it is carried down to the sea. It should be noted, however, that the movement near the bottom is always upstream, for example in Chesapeake Bay. Salt water does not penetrate far into estuaries: it reaches Quillebocuf on the Seine and Paimboeuf on the Loire. Only below these points is there a salt-water tide: above the current only moves freshwater. This has been called a dynamic tide (Francis-Boeuf). Salinity in the downstream section varies with the state of the tide, the regime of the river and the time of the year, as floods and low-water conditions depend on the last. The pH also varies in the same way. Tidal currents in estuaries are usually relatively strong: the maximum on the surface in the Tees, for example, being more than 3 knots on the flood at half-tide.

A *bore* is really neither a current nor a wave, but shares the qualities of both. This rapid rise in level with its associated breaking wave, which reaches a height of 5–6 m. in the Amazon, accompanies the flood tide in certain estuaries, and may cause serious damage. The bores in French estuaries have been greatly reduced by dredging, especially in the Seine.

Discharge currents are very powerful in certain channels and may reinforce either the ebb or flood tidal current. In Norway the Salström may reach speeds of 16 knots (?) when the ebb current is reinforced by meltwater (Rouch, III, p. 122). In the Tuamotu islands, discharge currents are very strong, 10–12 knots in the Hao channel, as they are fed by water flung into the lagoons by the waves breaking round the atolls. The inflowing tidal current may be eliminated for days in heavy storms and is never as strong as the ebb current (French Navigational Instructions).

Tidal and discharge currents affect the whole mass of water, both in estuaries and elsewhere. In the Moluccas they are often strong in straits at a depth of 1,000–2,000 m. (Kuenen, p. 67) but always decrease towards the bottom.

At the Golden Gate, San Fransisco, the velocity 1 m. above the bottom is only half the surface velocity. Measurements made by van Veen (pp. 45–7) in the Straits of Dover have shown a variable relationship between surface velocities and those at 15 cm. from the bottom. The maximum velocity is usually found several metres below the surface, and the rate of decrease is only rapid close to the bottom. The same is true of the Tees estuary (Francis-Boeuf, p. 173). Lesser has also confirmed these measurements and shown that the decrease in velocity follows a logarithmic curve.

Other currents of prime importance in marine hydrology are of little importance in the study of coasts, but may, it appears, be significant in the formation of submarine relief. The main oceanic currents, which usually result from density differences and regular winds such as the monsoons or trades, have a negligible effect on coasts when compared with the coastal drift set up by oblique waves. But a wind blowing in the opposite direction to the coastal drift caused by oblique waves may, depending on the relative strength of the two forces, be the dominant factor in longshore drift. In deep water, where the main currents are formed, there is a marked deflection due to the effect of the earth's rotation, according to Ekman's law. But in shallow coastal water such deflection is negligible at all speeds (Sverdrup, Johnson, and Fleming, p. 495). The wind need not be parallel to the coast to cause a coastal drift, but it must not blow at right angles to the coast. It will be more effective if blowing onshore rather than offshore, as water will pile up near the land (cf. p. 27) and friction with the earth's surface will not reduce the wind so much.

Currents are turbulent flows, as can be seen in estuaries (p. 102),

where there is a considerable load in suspension and where the whirls of different coloured water show up clearly. It is thought, however, that there may be a more laminar flow near the bottom (Kuenen, p. 255, Hjulström, p. 328).

General effects of currents on transport and deposition. We can only deal briefly with this problem, which, although of the greatest importance, is not directly related to our theme.

In calm water all suspended particles must be deposited. Theoretically, the settling velocity is proportional to the square of the radius of the particles provided they are spherical, i.e. $v = Cr^2$ where C is a constant depending on the material. A sphere of quartz of 1 μ diameter thus takes 24 hours to fall 10 cm. In practice, this formula applies only to particles no greater than 0·1 mm. in diameter. Above this limit the velocities are less than those given by the formula. The form of the particles has the greatest effect, however; flat particles fall very slowly as they remain horizontal.

Currents interfere with sedimentation either by eroding loose particles or by preventing them from settling. Either erosion or transport or deposition may, therefore, take place. The relations of these to the velocity of the current and to particle size are set out in the curves published by Hjulstrom (1935, pp. 297–8) (Fig. 1 H). The upper curve shows that the velocity needed to move loose material on the sea-floor decreases with particle size down to 0·25 mm. approximately. Below this size, the velocity increases again and the velocity required to move a particle of ·002 mm., 180 cm. per sec., is the same as that required to move particles of 20 mm. diameter. Thus, sand is most readily moved, while both silt and gravel require higher velocities. This is especially important in the geomorphology of estuaries. It should be noted, however, that the law is based upon sediments composed of one particle size only.

The minimum speed required for transport (lower curve) appears to decrease regularly down to the finest particle. In reality, the nature of the curve below 5 mm. is uncertain as this was the smallest size included in Schaffernak's measurements on which the curve is based (Hjulstrom, p. 320).

Transport by currents and by winds is effected in four ways: suspension, saltation, rolling, and dragging, as Gilbert (1914) demonstrated. The velocity of particles in suspension is equal to that of the currents. In rolling the velocity of the particles increases rapidly as soon as the velocity required to set them in motion is exceeded, but further increases

in velocity cause only slight increases in the velocity of the particle. There are also sudden increases of speed, at least for small pebbles (Blackwell in Hjulström, p. 336), when the motion changes to suspension. Saltation is aided by the turbulent layer above the zone of generally laminar flow and to a lesser extent by slight turbulence in the laminar layer itself. Particles are picked up, fall, and set other particles in motion.

C. SUBAERIAL RUN-OFF, INFILTRATION, AND FROST ACTION

The rocky areas near to and immediately above the shore may be compared with semi-arid regions, as salt foam and spray largely inhibit the growth of vegetation there. Thus, there is little vegetation to check the run-off. But, in contrast with semi-arid regions, the 'precipitation' is enormous, if we add the sea-water flung on them to the normal rainfall. The absence of vegetation and the 'precipitation' of as much as several dozen metres a year in some places, combine to create conditions which are not found outside the coastal zone.

All land water, however, does not run off. On many convex cliffs vegetation covers over half the height, the only bare area being near their base. In addition, when the cliffs are cut into a plateau, the plateau surface is usually covered with vegetation. If the rocks are fissured, water percolates along the joints. In limestones solution takes place along the joints, and the cliff may be turned into a karst. If permeable rocks overlie clays in the cliff, and provided that the rocks are bedded and dip towards the sea or are horizontal, a water table forms above the junction of the two beds and saturates the clays, a condition very favourable to landslips. An alternation of sands and shales, truncated by the top of the cliffs, may produce comparable though more complex effects, provided that there is a marked dip towards the sea.

Frost plays a part in the enlargement of fissures in the cliffs in cold and temperate regions. But its main effect is at the base of the cliffs, as has been shown by Nansen (1922, pp. 28–32) and other Scandinavian authors, especially Vogt. In cold seas, such as those of Norway and Greenland, an ice-foot or ledge of ice forms along the cliff about half a metre above high tide level and is not carried away by the ebb. It melts only partially in late summer. If this ice were formed from sea-water it would have no great power of expansion, as sea-ice includes interstitial salts thrown out during freezing and these render it cellular and relatively soft. But the ice is also formed partly of snow drifts and

to a considerable extent from the water of fjords, which has a low salinity because of the freshwater brought in by rivers, and finally from run off which freezes on it. Thus, the ice becomes almost as hard as freshwater ice. The number of freeze-thaw alternations is very large, as both diurnal heating and cooling and periodic splashing with salt water may cause them. The meltwater penetrates into cracks in the rock, freezes there, and shatters the rock a little above high tide level. All the waves have to do is to remove the shattered debris. The degree of fracturing of the rock is obviously of greater importance than its resistance to impact.

Ice-feet of different characters have been described by British authors in the Antarctic. Their chief effect is to deaden the breaking waves.

Gilbert observed another process connected with freezing. When small lakes freeze, the expansion of the ice tends to dislocate the cliffs, or if it acts on a beach, to produce a small rampart or shore-wall, beyond the limit of the waves.

English authors have often resorted to ice-rafts to explain erratics in raised beaches in Great Britain. Barrois in 1876 used the same idea to explain comparable features on the Brittany coast. In New England and Acadia, such ice-rafts at present add to sedimentation in coastal marshes. They are carried in at high tide and leave a layer of sediment when they melt (Johnson, 1926, pp. 589–90).

Considerable erosive power has sometimes been attributed to fragments of ice, either hurled against the cliffs like pebbles, or dragged along the coast by currents, thus enlarging certain estuaries (Johnson, 1926, pp. 590–1). Nansen (1922, p. 32 and p. 41) believed that these processes were not very effective, and discounted the idea that such lumps of ice could transport much material. But much must depend on local conditions, since the openness of bays must affect transport, and the nature of the rocks the amount of erosion.

Finally, because sea-level was much lower in the last glaciation, present coasts may have been affected by periglacial processes at that period. Thus, unless there has been much marine erosion since, we may find Pleistocene freeze-thaw effects such as rocks shattered by frost, old solifluxion forms and polygonal soils on the present coasts.

D. WIND

Wind action is very important but usually makes itself felt indirectly through waves and the spray which it beats against the cliffs (see above).

Further, the wind transports beach sand inland, where there is an abundant supply, to form coastal dunes. The mechanism of sand transport is the same as in inland dunes and will not be described here. Coastal dunes themselves are considered in Chapter III.

Violent storms and tropical hurricanes blowing onshore not only cause waves to act at higher levels than normal, but also give rise to variations of sea-level of 2–3 m. (Fig. 1 G) and even, it appears, of 5–6 m. in certain bays. These variations are known as storm surges. They are very important on flat coasts such as those of the North Sea, the Baltic, and the Gulf of Mexico, where such surges, if they coincide with high tide, may cause catastrophic floods by breaching sea-walls or natural sand and shingle forms, as happened on the 1st February 1953 in south-east England and the Low Countries. They may also occur on atolls. Constructional forms well above the level of normal storms do not necessarily imply former higher sea-levels, but may result from abnormally severe storms. On the other hand, a period of offshore winds may cause tides lower than can be explained by tidal action alone.

E. CHEMICAL PROCESSES

The effect of chemical processes on coastal evolution attracted attention much later than mechanical erosion. There is much still to be learned of their exact importance, but their effects seem to be established and are probably considerable.

(a) *Solution of limestones.* Many facts, which will be treated in Chapter III, lead to the belief that limestone coasts are subject to solution in both the intertidal and spray zones. Such solution would present no difficulty, if it had not been observed, at least in warm seas, that the sea is apparently saturated at the surface with calcium carbonate. This was observed by Murray a long time ago and by Vaughan, Wells, Revelle, and Wattenberg more recently. In places it has been shown that supersaturation occurs.

It has been suggested that the difficulty might be met by recourse to freshwater from the land, percolating into the limestone and causing solution down to the level permanently saturated by the sea. Solution above low tide level would thus be explained, as suggested by Wentworth in his study of Oahu in the Hawaiian Islands.

The hypothesis, however, meets three serious objections.

In the first place, it is very doubtful whether the amount of freshwater would exceed the amount of sea-water in the intertidal and spray

zones, even in a relatively wet climate. In Oahu, where corrosion is intense, the limestone regions only receive 625 mm. of rain a year, according to Wentworth.

Secondly, as observed by Panzer, corrosion forms at the foot of certain isolated rocks in the Palaos Islands in the Bismarck archipelago and elsewhere cannot be attributed to freshwater because the surface of the islets is insufficient to nourish either appreciable run off or underground drainage.

Finally, corrosion forms, as well developed as those in humid regions, are found in the Red Sea, an arid region where the action of freshwater may be discounted.

The explanation may partly lie in variations in the amount of carbon dioxide dissolved in coastal waters and especially in pools. These are reflected in the pH values. Davy de Virville has proved that green algae living in pools raise the pH during the day. Emery has shown that the carbon-dioxide content in solution basins in calcareous sandstone in California is higher during the night than the day, i.e. the pH value is lower. Such changes are caused by the nocturnal emission of carbon dioxide by algae and invertebrates. During the day, although the latter may still produce carbon dioxide, green algae at least absorb it in the process of photosynthesis. This abstraction of carbon dioxide from the water causes the deposition of finely divided calcium carbonate in the pools, as the water becomes saturated. But such calcium carbonate has not time to adhere to the bottom but adheres only to the ridges separating the pools, where it builds up as it is most frequently subject to drying. Elsewhere it is removed at high tide, so that solution takes place at night when carbon dioxide is emitted by the algae. The calcite cement of the sandstones is attacked, and the loosened sand grains removed by the waves or by periwinkles (*Littorina*), which are very common in California. Emery has shown that 2,600 periwinkles can remove 0·3 gm. of sand grains in 24 hours. The periwinkles thus play two parts, they produce a continual emission of carbon dioxide and also complete the work which solution begins. The sand grains ejected by the periwinkles are later removed by the waves at high tide.

The diurnal phenomena had been described before by Davy de Virville and by Sverdrup, Johnson, and Fleming who state that the precipitation of calcium carbonate is due to the abstraction of carbon dioxide during photosynthesis (p. 950). Emery's important contribution was the recognition of the opposite process at night. The solution process would be, therefore, biochemical.

At about the same time, Fairbridge (1946–7) came to similar conclusions in Australia. In addition, he mentioned the effect of nocturnal cooling, since carbon dioxide is more soluble in cold water than in warm water (Williams, 1949). But it seems doubtful whether cooling has any direct effect in coastal solution; and, more recently, Ranson has suggested that biological processes alone can account for coastal corrosion phenomena in limestones (see below, p. 29). However, experiments carried out by others (Guilcher and Pont, 1957) have shown that, in some calcareous rocks at least, cycles of wetting and drying, combined with solution, play a very important part in disintegration. In sum, the coastal corrosion of limestones seems to result from a number of processes.

(*b*) *Formation of coastal calcareous sandstone*. The problem is further complicated by the fact that corrosion may predominate on one part of the shore and formation of sandstone on another. Discontinuous beds of calcareous sandstone are very common on sandy beaches in coral regions where they are known as 'beach-rock' (cf. p. 122), but precipitation has also been observed in certain temperate regions, such as the coasts of Victoria (Hills, 1949).

Such sandstones in coral beaches have been explained in a number of ways: by evaporation of salt water contained in the sand (Saville Kent, Stanley Gardiner, 1902–6, and Daly); by freshwater percolating into the beach (David); and at the contact of freshwater and salt water in the sand (Stanley Gardiner, 1930). Sewell and Kuenen (1933, pp. 87–8) consider that there are two types of cementation: one occurs above high tide level and seems to be connected with freshwater action; the other, which results in several beds one above the other, seems to be the result of sea-water, as it is found in areas where there is no possibility of freshwater. The first process produces horizontal beds and the second beds parallel to the beach surface. Ginsburg states that beach sandstones in Florida are produced by precipitation of aragonite at low tide.

In a very detailed work, Bavendamm has shown that in vegetable muds in mangrove swamps in the Bahamas, bacteria are of the greatest importance in the precipitation of calcium carbonate. Many tropical marine limestones might be explained by bacteria; Nesteroff thinks that the beach-rock of coral regions is related to the metabolism of bacteria living in the sand, leading to precipitation of limestone. According to Ranson, the organic matter decomposed by bacteria would account for beach-rock formation, without precipitation of limestone at all.

On the other hand, Hills (pp. 146–7) has noticed a secondary cementation by calcite at the foot of cliffs in dune limestones in Victoria: this cementation is most marked at mean sea-level. He attributed it to the deposition of calcium carbonate from rain water percolating through the sand and meeting the zone saturated with sea-water.

(*c*) *Water layer weathering* has been studied in Australia, New Zealand, and the Hawaiian Islands. In some rocks, such as altered basalts, volcanic tuff, kaolinized granites, certain sandstones and shales, the edges of the pools, especially in the spray zone, are actively eroded. The pools are bordered by small overhangs and ultimately merge into each othe. Quartzites and parts impregnated with limonite resist erosion and form small-scale residual relief. In Victoria, in sandstones in which the joints are filled with limonite, pools form in the sandstone and are bordered by small limonite ramparts, a centimetre or so high. In the same area on Phillip Island, where the joints of columnar basalts have been altered, pools form along the joints and are separated by the remains of the basalt polygons. At Fingal's Cave and the Giant's Causeway, on the other hand, the pools form on the actual columns of basalt (Panzer, p. 33). English and American authors have attributed these effects to alternate wetting and drying, which is especially characteristic of the spray zone, where the pools may be filled in storms and partly evaporated in calm weather. Bourcart attributes a major role in honeycomb weathering to alkaline chlorides brought up by water but the forms which he describes are not entirely coastal.

The experiments of J. Joly have shown that basalt, obsidian, hornblende, and orthoclase are 3–14 times more soluble in salt water than in freshwater. Solution may, therefore, affect rocks other than limestones. Further, forms like those on the edges of pools in the Pacific also occur on limestones of varying age and hardness in Ireland, Wales, Morocco, Provence, and the Alpes-Maritimes (cf. pp. 65–6). Such erosion, which extends below the level of high neap tides, appears to be better explained simply by the wetting of the rocks than by the number of cycles of wetting and drying.

In view of changes of sea-level, we must not forget the possibility of former subaerial chemical weathering on present shorelines. If, for example, a coast had been subjected to tropical weathering, the rocks may have been deeply rotted, with the formation of laterite. There are also a number of drowned karst regions. We must bear it in mind along with the possibility of Pleistocene periglacial effects.

F. BIOLOGICAL FACTORS

Corals and calcareous algae form reefs, which are the most widespread form of coastal relief in warm seas. We can scarcely separate the study of the ways in which they act from the study of the forms they produce (Chapter III).

Soft algae, unlike calcareous algae, are especially well developed in temperate seas, where they may reach lengths of over 10 m. They are important because of the braking effect which some of them exert on the waves. On the rocky coasts of Great Britain and Brittany, there is a zonal division of vegetation on the beach. The most important zones from our point of view are the following:

(*a*) The highest is the *Fucus* zone, with *Ascophyllum nodosum* in its upper parts.

(*b*) The zone of *Himanthalia lorea*.

(*c*) The zone of *Laminaria* which extends below low tide level.

Fucus exerts little braking effect on the waves as it cannot grow where the breakers are strong; on the island of Sein, it occurs only in relatively sheltered localities, such as flat areas and especially hollows.

On the other hand *Himanthalia lorea*, *Laminaria saccharina*, *L. flexicaulis*, *L. cloustoni*, and *Saccorhiza bulbosa* can put up with breakers. When they are fully grown, these plants, which are about 3 m long, exert a marked braking effect on the surface of the sea, when they are exposed at low tide. Provided that the waves are comparatively small and that the offshore slope is not too steep, these algae completely stop the waves from breaking, until the tide is high enough for them to be submerged. This action, together with the effect of the beach profile (p. 18), helps to explain the differences in wave size between high and low tides.

Further, large algae attached to stones may be flung up the beach in storms and carry the stones with them. Some invertebrates, such as *Pholas* and *Patella* and *Cliona*, bore into certain rocks, especially limestone, and convert them into sponge-like forms. Their effects have been studied mainly in coral regions; some bore by mechanical methods and others by secreting acids. The holes they make extend the surface open to the attack of other agents, and this is probably their most important function. Finally, burrowing organisms may reduce sediment to powder by passing it through their digestive apparatus. In tropical seas, an enormous amount of sand passes through *Holothurians*, although these do not comminute the sand except perhaps very slowly.

But they are not the only creatures concerned. Ginsburg goes so far as to explain micro-karst forms in the intertidal zone, which are usually attributed to solution, by the action of boring and burrowing organisms: but according to Nadson, the holes made in limestone by algae are a form of solution. Ranson goes so far as to explain the coastal corrosion of limestones, entirely by the action of microscopic algae (Cyanophyceae). However, although he supposes, as Nadson does, that this action is chemical rather than mechanical, his opinion seems to be too extreme.

Although observations are scarce, it appears that fish, crustacea, and mollusca help to open up fissured rocks: congers, crabs, and certain gastropods insinuate themselves into fissures; crabs attacked in their holes brace themselves against the rock and thus act as levers. Certain fish in the Red Sea eat coral. A detailed examination of that part of the shore, which is uncovered by the sea only on rare occasions, should be interesting, provided that it is made with care, for most of the loosening of rocks in this zone is the work of fishermen.

Sedimentation is caused on the coast of Provence by banks of *Posidonia*: these sea-weeds can produce a rise in the sea-floor of 1 m. in 100 years, as they cause particles to be deposited.

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Chapter II

SHORELINE MOVEMENTS

A. THE DIFFICULTIES OF THE STUDY

The problem of shoreline movements is one of the most important and at the same time most difficult aspects of geomorphology. The enormous growing bibliography merely increases the confusion. Research methods are complicated and critical, and part of the divergence of opinion results from research incorrectly carried out.

In the reconstruction of former shorelines, we may rely either on emerged or drowned coast forms, or on shallow-water deposits no longer found at the edge of the sea, or, and this is best though rarely practicable, on evidence of both types.

Deposits can only betray the position of a former shoreline without question when they take the form of a shingle beach, in which case they indicate the former high tide level except in the rare instance where the whole of the beach is composed of shingle, as at Chesil beach in Dorset. Even coarse sand can be deposited at considerable depth. Solitary pebbles may be deposited on the floor of the sea by floating ice, as has perhaps happened off the Brittany coast; they may be deposited on land by solifluxion or some other process, provided they do not occur in hollows beneath an overhang. An isolated shingle beach does not necessarily imply a period of stability: it may have been formed during either a slow transgression or a slow regression and have remained where it is for some local reason.

Among landforms dead cliffs are valuable though not indispensable evidence, as all shores are not cliffed. But every steep slope behind flat surfaces is not necessarily a sea-cliff, even if it is vertical. If it has caves with well-rounded shingle and boulders at its base, the case is usually proved, but some doubt must remain concerning granite boulders, such as those of the Cadillac Cliffs (New England) which may well not be marine. We must also guard against the resemblance between sea-cliffs and the steep slopes behind pediments. Further, dead cliffs are

often quickly degraded, as are those of the Bas-Champs of Picardy: their base is buried under and smoothed out by solifluxion or material brought down by creep, which masks the old beaches except where a quarry happens to reveal them. In general, a constant elevation over a considerable distance at the foot of the cliff is evidence in favour of a marine origin: but the opposite case does not mean that we can discount the sea as the agent of formation, as the old shoreline may have been warped.

D. W. Johnson has shown that every shoreline slightly above the level now reached by the waves is not necessarily a proof of a general change of level. In bays with narrow entrances, such as the Etel river or Morbihan, high tide level is slightly lower than outside the bay. If the narrow entrance should be widened, the tidal range increases and high water mark rises: the opposite may happen if the entrance narrows as in the *graus*, or narrow openings into lagoons in southern France. On the other hand the level of storm beaches is not everywhere the same, even outside sheltered bays. It depends on the exposure to strong waves, the local tide range, and the amount of rise in sea-level occurring in storms (p. 27), so that beach ridges formed under present conditions may have levels which differ from each other by as much as several metres. Spray may also form small solution terraces above present sea-level. It may also happen, even in localities of identical exposure, that two contemporaneous shingle ridges are not at the same height above present high tide level, because the former high tide level which they represent was not at the same height above the present level in both places. We know that every tide results from a superposition of waves variously affected by resonance phenomena, which are a function of the form of the basins concerned. Thus, when sea-level was higher, the tidal range may have been quite different locally because the form of the basin was not the same.

These observations indicate how useless it is to try to establish the altitudes of old shorelines in tidal seas to the nearest metre by means of deposits. Much greater accuracy can be obtained by studying the solution grooves formed on limestones in warm seas with very little tidal range.

The levels of terraces or river flats are sometimes used to establish old high sea-levels: this has been done for the Loire by Chaput and for the small streams of the Channel Islands by Hanson-Lowe. This method is valid when the river terraces grade into marine terraces near the mouth. If they do not, there is considerable uncertainty as to the heights

of old shorelines as their positions are not known where should the long profile reconstructed from the terraces be stopped?

Pre-Pliocene shorelines are rarely of geomorphological interest. If they are very old they may be obliterated, while the older they are the greater the chance that they are deformed. More forms have survived from the Pliocene. In southern England the dip-slopes have been bevelled by platforms, which are certainly marine and of Pliocene age. The most vexed of Quaternary problems begins in the Pliocene: are recent movements of the shorelines due principally to tectonic movements or to variations in sea-level?

B. EUSTATIC MOVEMENTS

Shoreline changes caused by changes in sea-level are termed eustatic. There are two types of eustatism. The first is diastrophic and is connected with changes in the form and depth of the oceans. Suess and de Lamothe both stressed this, but it has been invoked most insistently and most clearly demonstrated by Baulig in accounting for Pliocene planations in various regions. The second is glacial and is caused by changes in the volume of the continental ice sheets; as a result, it affected only the Quaternary period. In effect, every variation in the amount of ice caused an opposite variation of sea-level, because the water involved must have been derived from the sea via the atmosphere. As it is known that there were several glaciations separated by interglacial periods, although the exact number is still open to question, we must expect each glacial period to have been associated with a eustatic fall and each interglacial with a eustatic rise of sea-level.[1]

Thus there evolved, largely on Depéret's authority, the idea of a series of Quaternary marine terraces. The idea was partly based on the argument outlined above but also on forms and deposits found at constant elevations, which appeared to verify the hypothesis. It may be said that no idea had quite such a vogue as this in historical geomorphology; it may be readily summarized (Fig. 2). In the Quaternary the sea occupied four levels: 80–100 m (Sicilian), 55–60 m (Milazzian), 30–35 m (Tyrrhenian), and 15–20 m (Monastirian). Recently various authors have stressed the importance of a fifth level, 2–5 m above present high tide level. Various names have been proposed for this: Normannian (by Dangeard in the Armorican massif), Lower Monastirian (Guilcher in Brittany), and Ouljian (Gigout in Morocco). The lower the level the more recent it is. The Monastirian cannot be

Brittany, Normandy, Jersey, Guernsey, Portugal, the Azores, and Morocco. In many places it survives in most coves, but is occasionally continuous over distances of some kilometres. Its equivalent has been reported from as far afield as Australia. Usually, this shoreline is better represented than the Monastirian proper. It has been interpreted as a stage in the retreat from the Monastirian level of greater duration than that responsible for the Monastirian itself. It is clearly earlier than the last glaciation, the deposits of which cover it. In the whole of north-west Europe it has been affected by periglacial action.

Two other stages evidenced everywhere are the pre-Flandrian regression and the Flandrian transgression. These have been studied in great detail in the coastal plains of north-west Europe, on the Thames, the Lower Rhône, and the plains of Italy bordering the Tyrrhenian Sea. They are evidenced in many other regions by the drowning of the lower parts of valleys, the submergence of peats and megalithic monuments and the deposition of muds. It is generally but not universally agreed that the pre-Flandrian regression reached a level of 100 m. But one of the earlier regressions, the pre-Monastirian or pre-Tyrrhenian, may have been even lower as is suggested by ridges of pebbles off the west coast of Brittany. Evidence in favour of the idea that the maximum of the Flandrian transgression (the Dunkirkian) was 2–3 m higher than the present level is provided by grooves or notches slightly above present high tide level in various coral regions, and by post-Flandrian deposits, some of them accurately dated and not appearing to be the result of storms affecting present sea-level, south of the bay of Audierne in Brittany and around Rabat.

C. QUATERNARY TECTONIC AND EPEIROGENIC MOVEMENTS

Quaternary shoreline movements have also been attributed to earth movements, either isostatic and resulting from the glaciations, or independent of the glaciations.

Glacial isostasy is generally accepted and only the details need be discussed. The loading of the ice-sheets caused a plastic depression of the earth's crust and perhaps a slight complementary elevation near the edge of the ice. When the ice-sheets melted the land formerly covered by them rose again, while there was perhaps a depression at the edges. These movements were relatively slow. In effect, the prolonged overloading caused, in the earth's interior, a slow flow of viscous material

towards the periphery of the areas affected to re-establish an equilibrium. The removal of the load caused an equally slow transfer of material in the opposite direction.

The existence of these movements has been beautifully demonstrated in the area of the Fenno-Scandian ice-sheet by the study of the present altitudes of raised beaches (Fig 2 B). The isobases, or contours of the elevations to which the various beaches have been uplifted, show no uplift in the south of the Baltic The amount of uplift increases to a point in the north-west of the Gulf of Bothnia, where the ice was thickest. The same studies have revealed further evidence of the Flandrian transgression which was superimposed on the isostatic uplift. The Baltic was a lake about 9000 B C when the ice-front was in the vicinity of Lake Vanern in Sweden It then became a sea, the Yoldia Sea, in about 7900 B C after the retreat of the ice had opened the way for a marine transgression across central Sweden before isostatic uplift really set in The isobases of the associated beach reach 275 m in the north-west of the Gulf of Bothnia Later, isostatic uplift isolated the Baltic which became the Ancylus Lake about 6500 B C the isobases of the associated beach conform to those of the Yoldia Sea, but only reach 180 m Later, about 5000 B C , the continuation of the marine transgression opened the strait between Denmark and Sweden in the peripheral region where isostatic uplift was very slight The following phase was that of the Littorina Sea, with a maximum isobase of 120 m The uplift still continues, the maximum value, 10 mm per year, occurring at the north of the Gulf of Bothnia

Similar phenomena occurred in all coastal regions occupied by ice-sheets On the north-east coast of North America, the uplift increases west from a line drawn between New York and the south of Newfoundland, which is effectively the zero isobase It increases towards the St Lawrence estuary and on the north bank of that river reaches a value of 150 m (Flint) The uplift has not ceased there In Scotland beaches have been uplifted isostatically to 30 m , the amount decreasing southwards to nil in South Wales In the polar regions, the data are less precise, but the uplift was doubtless considerable The Russians have observed raised beaches up to 400 m in Novaya Zemlya, where a series of beaches occur on the north coast below this level In the Kola peninsula the uplift appears to have reached 230 m , while in Franz-Josef Land it reaches 333 m The shrinking of the Antarctic ice-sheet has had the same effect, but the amount is uncertain

Isostatic uplift after the Ice Age has thus brought into being series

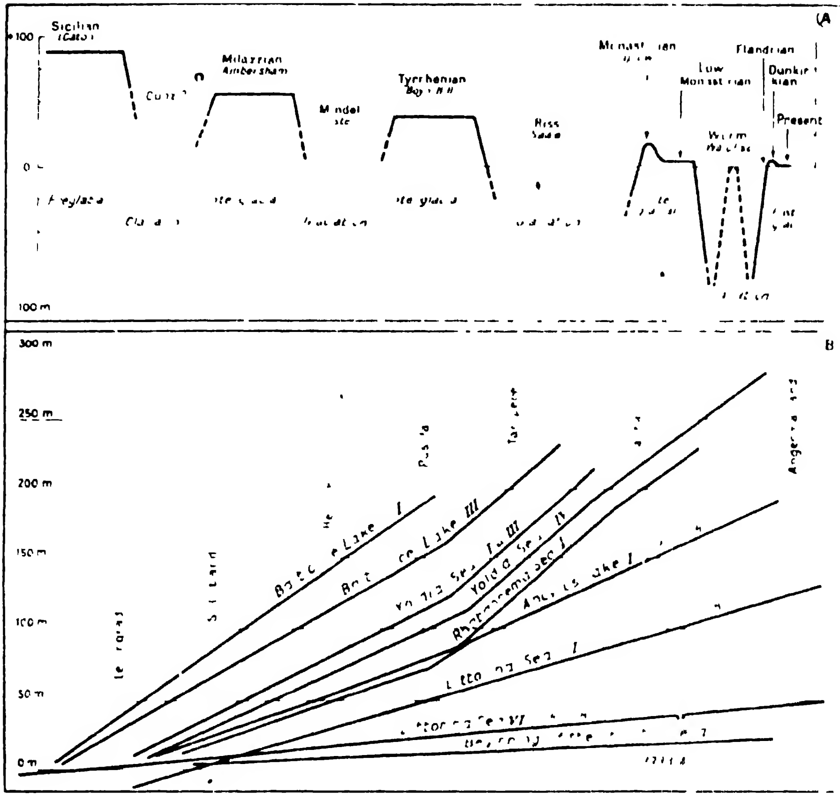

FIG. 2 EUSTATISM AND ISOSTASY

A. Diagrammatic representation of glacio-eustatic variations in the Pleistocene. The interglacial names in brackets are those used by Bull (1942). The names of glaciations in brackets are those applied to North Germany. B. Simplified diagram of glacio-eustatic uplift in the Baltic (after Sauramo, 1940). Altitudes in metres are above the 1938 sea-level The horizontal represents the area between the point of maximum uplift on the Angermanland coast in Sweden south and south-eastwards towards Leningrad. The curves show the uplift of successive shorelines in different places, the oldest being the highest. It will be seen that the uplift has not been uniform.

of beaches of much fresher appearance and locally more numerous than those due to eustatic movements.

Tectonic movements in unglaciated regions Some authors, while recognizing the existence of eustatic movements, have considered tectonic movements to be more important in the Quaternary even in extra-glacial regions. They attribute uplift of shorelines to such movements Thus, Fabre thinks that the deposits of Medoc and those beneath the sand in the north of the Landes in Gascony are the result of Quaternary earth movements But Bourcart in France and Jessen in Germany are the chief protagonists of the hypothesis of extra-glacial tectonic movements in Holland, Umbgrove appears to favour Bourcart's ideas

Bourcart and Jessen believe that the edges of continents are in special positions in relation to the earth's crust they are said to be subject to periodic expansion, the cause of which remains to be discovered, as the suggested explanations are only hypotheses These expansions alternate with contractions The expansions give the illusion of marine regressions, while the contractions simulate transgressions of the sea On the whole expansion dominates on the continents hence the facts that the higher terraces are the older ones, and the higher ones are farthest inland The same hypothesis also explains the fact that the edges of the continents are often zones of uplift, folds, horsts, or anticlinoria, while their centres are relatively depressed One of Jessen's maps shows this feature well it is fundamental in North and South America, Africa, Arabia, and India There should be, therefore, at the edge of the continents a marginal flexure, which would progressively deepen the seas near the continents as the edges of the continents were elevated Fig 33 D

This theory implies an upward deformation of the land and a depression of the sea-floor We shall see in the second part of this book the deductions that can be made from a study of the continental shelf, and we will concentrate at present on uplifted shorelines Bourcart maintains that the old, uplifted Quaternary shorelines of the coast of Morocco are appreciably warped He also makes a point of the intersection of terraces at the mouth of the Bou Regreg, the older ones being highest in the interior and youngest highest near the coast He thinks that the flexure is present on all European coasts and has sought evidence in the attitude of Portuguese, Breton, and Provençal Quaternary deposits Like many others he stresses the arbitrary and artificial character of

many eustatic interpretations: the facile correlation of terraces which are not really at the same altitude, the use of deposits of doubtful value, and insufficient morphological evidence.

He has met, however, considerable opposition from various quarters. The pioneer work he did on the Moroccan coast has not always been confirmed by later and more detailed work. The marine Quaternary deposits at Cape Cantin, Cape Tafelneh, Cape Rhir, at Agadir, and in south Morocco do not generally show warping. The higher marine deposits are not Quaternary but Pliocene, and deposits of that age are admittedly warped. In Morocco the classic eustatic theory seems to account for the large majority of the facts. It can even explain the intersecting terraces of the Bou Regreg, if we admit that the lower parts of the older terraces were deposited as the sea retreated to lower levels: their slope, 3 degrees at the junction with the underlying Miocene, is far from being unknown on modern beaches (p. 80). In Portugal, Zbyszewski, Breuil, and Teixeira have also adopted the eustatic hypothesis and it seems to us to be the best explanation of the terraces, at least near Lisbon. In Brittany, the presence of Lower Monastirian beaches at the heads of certain rias, such as Quimper and the Gulf of Morbihan, at the same altitude as on the coast, is evidence against the idea that there has been warping since the beaches were formed. Such warping would have uplifted those parts of the beach nearest the interior of the land.

Must we then abandon the hypothesis of a marginal flexure? Certainly not, although its application should be restricted. It is possible that there was a post-Tyrrhenian flexure in the Hyères islands in Provence, as thought by Bourcart: his interpretation of the section of the Pointe du Tuf at Port Cros is justifiable. In eastern Corsica, some support can be found for the hypothesis. In Africa proper, i.e. well away from the Alpine region, it provides a satisfactory explanation for many of the shorelines. J. Dresch resorts to it, and it is significant that Jessen's theory was largely based on the coastal uplift of Angola.

C. Arambourg and A. Cailleux have, on the other hand, recently put forward the idea that erosion of material from the land causes an isostatic uplift of old shorelines, most marked in the oldest shorelines. Such an uplift is possible, but, as H. Baulig observed and as Arambourg himself recognized, Quaternary erosion must have varied greatly from place to place, so that it is difficult to believe that it could have resulted in uniform uplift. In addition, isostatic readjustment does not seem to follow small changes of load. Finally, a coast such as the Atlantic coast

of Morocco was not eroded but loaded in the Quaternary period with great masses of dunes which were quickly consolidated. Yet the classic series of terraces is well developed there, a fact which casts doubt on the value of the theory

We may come to the following conclusions The Quaternary, which is of greater interest than earlier periods to coastal geomorphology simply because it is the most recent, was not a period of no earth movement. Thus we get rid of the anomaly, said to be incredible by Bourcart, of a geological period with no earth movements But, in view of its recent age and its shortness, it is quite in order to attribute little deformation to it It is unique, as a result of the many eustatic movements attendant upon the glaciations on the whole these were more important than tectonic movements There is no need to abandon Deperet's system of Quaternary terraces Further research has indeed shown in many places the weaknesses of the hypotheses of many upholders of the eustatic theory for example, practically nothing of de Lamothe's work on the Sahel of Algiers can stand after the work of F Glangeaud But in many other places, the existence of a series of unwarped terraces at 80–100, 55–60, 30–35, 20–15, and 5 m has been confirmed The most economical hypothesis to explain this uniformity is the eustatic one There is no compelling reason to abandon diastrophic eustatism in the Pliocene, but warping of shorelines should be superimposed on eustatic effects in more places than in the Quaternary In Flanders the Lower Pliocene platform has been warped down towards the Low Countries In short, the further one goes back, the greater the chance of tectonic movement We must naturally make an exception of the regions affected by Quaternary ice-sheets, where warping of a special type has certainly occurred and still continues

Zenkovitch has drawn attention to other aspects of the formation of marine abrasion surfaces If the amount of material brought down by the rivers has varied with climatic changes, such as glaciation, the foot of the cliff may have been fossilized during phases when much material was brought in When the land rises steadily, one might have discontinuous abrasion platforms formed in periods when the rivers were supplying the sea with little material Further, alterations in wind regime, causing a variation in the strength and direction of the waves, affect the rate of formation of coastal platforms Similarly, changes in the area and shape of the sea-basin would have some effect

Whatever the exact truth, the past history of coasts is of the greatest importance in understanding the present To understand the form and

stage of evolution of the world's shorelines, we must remember that the sea has only been at its present level for a few thousand years. This is especially important on coasts which develop slowly and which owe much of their evolution, as a result, to the past. The morphology of coral reefs is also clarified by the study of eustatic and diastrophic movements. Because much of the past history of coasts is to be found in the beach deposits fringing them, the study of such residual deposits is as essential as the study of residual deposits in the interpretation of subaerial erosion surfaces.

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Chapter III

COASTAL FEATURES RELATED TO SEA ACTION

The problem of the classification of coasts. The problem of the classification of coasts is a difficult one and the discussion will be limited to the classifications proposed by Gulliver, Johnson, and Shepard.

Gulliver, in 1899, distinguished initial forms from sequential forms. Initial forms are not connected with the action of the sea as such, but are the result of relative movements of land and ocean. Lithological and tectonic factors, such as vulcanicity, and climatic factors, such as glaciation, may also produce initial forms, which Gulliver terms 'accidental'. Sequential forms evolve from the action of the sea on the initial forms.

Johnson in 1919 divided shorelines into four categories: shorelines of submergence, shorelines of emergence, neutral shorelines, and compound shorelines. Where submergence and emergence have occurred recently in the same place, the shoreline is attributed to that movement which has been responsible for producing its salient characteristics. The classification is, therefore, genetic. For each type, the author pictures an evolution of the shoreline under the action of the sea, so that he introduces, in effect, Gulliver's concept of sequential forms. Six years later, in his work on the coasts of New England and Acadia, Johnson reverted to Gulliver's classification with the following general plan:

I. Initial forms are divided into those:
  (*a*) Bordering high lands of hard rocks.
  (*b*) Bordering low lands of soft rocks.
  (*c*) Caused by glaciation.
  (*d*) Modified by slight variations of level.

II. Sequential forms are divided into those:
  (*a*) Bordering high lands of hard rocks.
  (*b*) Bordering low lands of soft rocks.

(*c*) Bordering areas of unconsolidated deposits.

(*d*) Consisting of coastal marshes.

In 1948 Shepard evolved a classification recalling in principle that of Gulliver. He also has two main types, initial and sequential, which he terms primary and secondary. The primary coasts include the forms due to submergence of a subaerially eroded land surface, e.g. rias and fjords; those formed of fluvial sediments, e.g. deltas and alluvial plains; those formed of glacial sediments; those due to the action of the wind, e.g. dunes and to the action of vegetation, e.g. mangroves; coasts associated with volcanic activity and with diastrophism. The secondary coasts are divided into coasts formed by marine erosion, and coasts of marine sedimentation, including coral reefs.

Thus, Gulliver's classification has so far dominated much of the literature in coasts. The reasons for abandoning Johnson's first classification are important, as a genetic classification is very tempting. But how are we to classify coasts which are still imperfectly known? New discoveries may cause descriptions, based upon principles subsequently recognized as false, to be abandoned. The general post-glacial rise in sea-level, commonly known as the Flandrian transgression, is a well-known phenomenon. All the coasts of the world are, therefore, coasts of submergence, except where very recent tectonic uplift, of an amount greater than the eustatic rise of sea-level, has occurred. Exception must also be made of coasts affected by post-glacial isostatic upwarping.

We will, therefore, adopt Gulliver's general principle. Although it may appear to be putting the cart before the horse, we propose to describe the sequential forms first. If we know the coastal forms which may be produced by the action of the sea, we shall be in a better position to understand the ways in which initial or primary forms may evolve. To argue from imagined primary forms to secondary forms involves the risk of too much speculation. No excuse is needed to emphasize the sequential forms for they alone are truly marine.

These forms can be grouped into four main divisions: cliffs and rock platforms; beaches and dunes; estuaries, marshes, and deltas; and coral formations. In the first group erosion predominates and in the other three accumulation. The division is not absolute for every beach undergoes occasional erosion; certain beaches bordering low-lying land are constantly retreating; at any one time the front of a delta is advancing in some places and being eroded in others; on the other hand, the slumping of a cliff may temporarily delay marine erosion by spreading

a mass of debris on the shore. Coastal dunes, estuaries, and deltas are classed as primary forms by Shepard but included here because the sea affects the nature of the vegetation and provides the material for dunes: it also provides in conjunction with rivers the environment where estuaries and deltas are formed.

A. ROCKY CLIFFS AND PLATFORMS

A cliff is a break of slope not covered by vegetation; its slope is usually steep, between about 15° and the vertical, or even overhanging: its height is very variable: it is found at the contact of land and sea, and its existence is usually due in some way to the sea. Landforms presenting more or less analogous characteristics but not associated with the sea may be termed false cliffs. The cliff usually has at its foot a rock platform of a much gentler slope than the cliff itself; this is traditionally known as an abrasion platform. But some cliffs, both true and false, plunge straight under the sea with no break of slope especially at headlands, e.g. Pointe du Raz, and the outer coast of Belle-Ile; such cliffs are called plunging cliffs in English. In bays cliffs are usually fronted by a beach instead of by a platform of solid rock. A rocky coast and a coast with cliffs are not necessarily synonymous: at Penmarc'h in Brittany and in certain places on the north-east coast of the Maures in Provence the coast is rocky without any trace of a cliff, even a false one; more common, on the other hand, as Bourcart emphasizes, are cliffs in non-consolidated wet sand, such as dunes or beaches in course of erosion, and in glacial deposits and in mud.

The classic theory of the formation of the cliff and of the rock platform which fronts it is very simple and it will be sufficient to mention it briefly (Fig. 3 A): the seaward steepening of the edge of the land results from the removal of a wedge of material by the mechanical action of the waves: the rock platform marks the level at which this action has stopped, at least temporarily; it continues to be worn down by waves loaded with material derived from the erosion of the cliff. Waves undermine the cliff and periodical collapse of the overhanging parts results. Similarly waves erode caves, the roofs of which may collapse. This process may be greatly helped by the compression of air trapped in them. Caves tend to be formed along lines of weakness such as non-resistant strata and faults. The abrasion platform is smoother than the cliffs, although in detail it shows many signs of structural control.

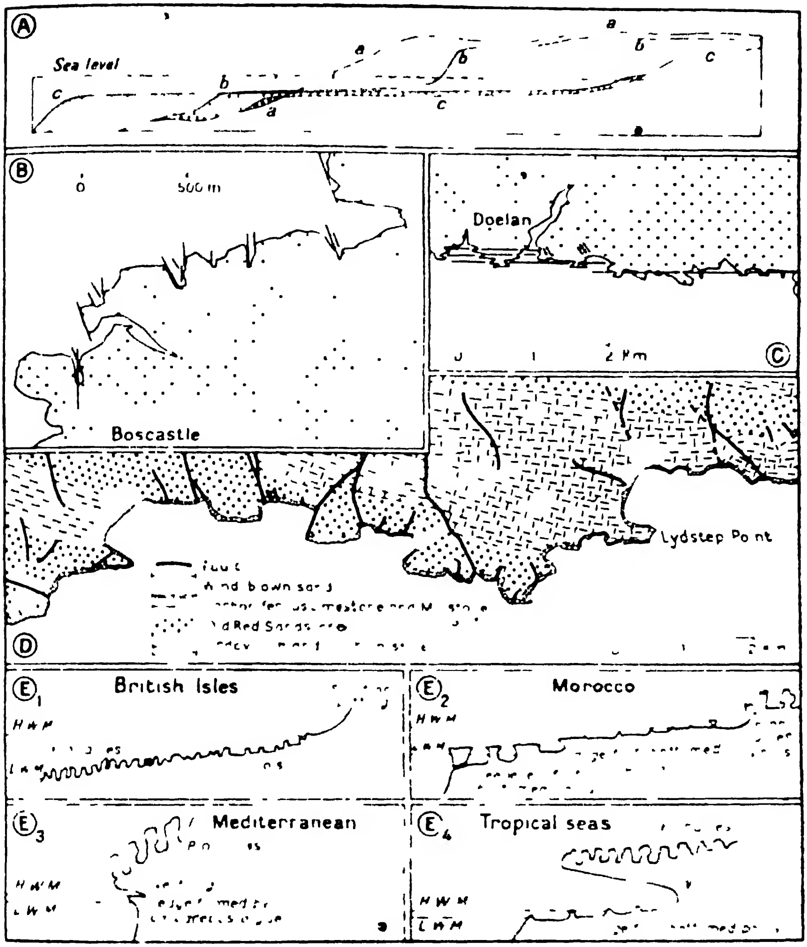


FIG. 3 CLIFFS: TYPES, STRUCTURAL EFFECTS, AND EVOLUTION

A. Classic concept of cliff recession with the development of a marine abrasion platform and the construction of a submarine terrace (after Johnson, 1919) on a submerged coast with a stable sea-level. B. Effect of joints on the North Cornwall coast (after Wilson, 1952). C. Effect of Palaeozoic faults on the Brittany coast east of Concarneau (after Guilcher, 1948): the main faults are east–west and the minor ones north-west–south-east. D. Effect of faults on the South Pembroke Cliffs (after 1″ Geological Survey map, No. 245). E¹–E⁴. Zoning of corrosion forms in limestones in different regions (after Guilcher, 1953).

This classic theory is applicable in certain cases, especially in connexion with the formation of sea-caves. It is undeniable that the sea hollows out caves by mechanical action. Conditions favourable for the formation of such caves are almost infinite in number. Basic veins in granite or metamorphic rocks are frequently etched out, e.g. on the Costa Brava in Catalonia, and in the region of St Malo and Dinard in Brittany. Bedding planes and lithological differences are often brought into relief, as in the Lower Devonian slates and quartzites at Pointe Armorique in the Rade de Brest, in the Carboniferous Limestone of the south coast of Pembrokeshire and the Gower peninsula, in the Old Red Sandstone of north-east Scotland and the Orkneys, in the limestone of the Gaspé peninsula in Acadia, and in the columnar basalt of Fingal's Cave. The most spectacular effects of the sea are to be seen in such forms. Apparently insignificant faults are likely to give rise to significant features in places swept by waves: this is clearly seen in the crystalline rocks of Sein island in Brittany, where long corridors are formed, in the metamorphic coast between Concarneau and Lorient, and in some sedimentary rocks, e.g. Carboniferous Limestone in Pembrokeshire and the Isle of Thanet Chalk. Clear examples have been described by Wilson from Tintagel in Cornwall (Fig. 3 B, C, and D). Usually the results are fissures several hundred metres or even more than a kilometre long, and caves of greater size than those caused simply by lithological differences. The caves often have narrow mouths but broaden out within. Collapse of the roofs is frequent, and gives rise to blow-holes which may resemble the swallow-holes and dolines of karst scenery, depending on whether they are complete or only partly formed as on the west coast of Belle-Ile. Naturally the form of the caves depends partly upon the dip of the beds or on the slope of the fault planes. As stated by Panzer, near vertical dips are the most favourable, but horizontal fissures may sometimes give rise to fine, flat-roofed caves as in Cape Aguilar at Hong Kong. Two sets of joint planes at right angles to one another are very favourable as in the Old Red Sandstone of Orkney. Either in the caves or at their entrance are found the largest rounded and polished boulders: blocks up to a cubic metre in size are not exceptional. They are evidence of the effectiveness of mechanical erosion in such places.

That notches or nips at the base of cliffs are due to mechanical erosion is often much more improbable. This explanation need not, however, be rejected in every case, notably in the dark friable Devonian shales of the Crozon peninsula, and the mica-schists between le Conquet and

St Mathieu point in Finistère. But where this groove or notch extends for a considerable distance, it appears to be usually due to chemical action. In this connexion we must distinguish between limestone and other rocks, and also envisage the evolution of the cliff and the abrasion platform as a whole.

A comprehensive examination of the known facts about limestones seems to show that there is a general zoning of forms between the area reached only by spray and low water mark. This zoning seems to be connected, perhaps indirectly, with the temperature, the forms being more developed in warm seas (Fig. 3 11–14).

(*a*) In cold and temperate seas, for example in the Carboniferous Limestone of the British Isles, one notices corrosion hollows a few millimetres in diameter forming a network all over the rocks in the spray zone. Lower down, in the area covered at high tide, these are replaced by pools with overhanging edges and flattish floors cut out of the rock. Lower still, as far as low water mark, networks of sharp pinnacles (coastal lapies) of considerable size are found in addition to the pools described above (pl. I B). These are much more jagged than those which occur in neighbouring limestones subjected to subaerial erosion. These coastal lapies have also been observed on the edge of lakes, notably in pre-Cambrian limestone in the Laurentian Shield (Aubert de la Rue) and in Carboniferous Limestone at Lough Erne in Ireland; therefore, salt water is not essential for their development. In western Crimea, Givago has noticed the same thing in Miocene limestone, but mechanical erosion has played some part here; soundings have shown that below sea-level solution forms were less important and much less characteristic.

(*b*) In warm waters, such as those of the coast of Atlantic Morocco, the zone of jagged lapies is higher up the beach in the zone covered by high tides and by spray. These lapies are riddled by corrosion and are separated by small flat-floored pools with overhanging edges. Lower down, midway between high and low water mark, they are replaced by a system of large flat-bottomed pools (pl. I C), separated by sinuous crests on the average several centimetres high and cutting the limestone platform into a series of steps. These crests are not formed of precipitated calcium carbonate, but are the remains of coastal lapies, as is shown by the presence of transition forms and the nature of the rock. Still lower down there is a zone where the pools are deepened to several decimetres, instead of centimetres, and the overhanging edges are submerged. The whole system ends at low water mark in a small vertical

or overhanging cliff. This has been observed in Morocco in limestones of diverse ages, Quaternary, Cretaceous, and Jurassic, all of which resist mechanical erosion.

(*c*) A rather different example is provided by the shores of Provence and the Estartit Massif in Catalonia, where below the lapiés, which are common in the zone of breaking waves and spray, there occurs an overhanging projection, generally less than a metre high, above a ledge formed of the calcareous alga *Tenarea tortuosa*. The constancy of the projection in this warm sea seems to be due to the small tidal range which ensures that solution is concentrated in a very narrow zone. Such a ledge is not confined to limestone coasts.

(*d*) In very warm seas, such as the coral seas, a zoning, like (*b*) above, also exists, notably at Oahu in the Hawaiian Islands and in the Red Sea; but in addition a new form appears, a considerable overhang at high tide level, termed the visor by Wentworth (pl. I D). This lies between the zone of lapiés, which is here even higher up the beach, and the rock platform with its system of large pools. Its base is at high tide level. The overhang in places projects more than 2 m. It occurs in both very sheltered places and on exposed coasts. It is found very frequently and often continuously for long distances in the Red Sea, in Bermuda, in Fiji, on the west coast of Luzon in the Philippines, in the Bismarck archipelago, in the Palaos, in Hawaii, in Western Australia, in Madagascar and in Tongking. In places it completely surrounds the base of small rocks which in this way develop into mushroom rocks. It is developed principally but not exclusively on raised coral reefs.

One must beware of attributing all coastal limestone forms to solution or some other kind of chemical corrosion. An examination of chalk ledges on the Normandy-Picardy border in the region of Tréport and Ault shows that boring made by limpets, mussels, and other organisms are very important, while mechanical action cannot be excluded, because of the lack of resistance of chalk to non-chemical processes. Pot-holes even occur in hard limestones in Morocco and in Oahu; in the Crimea mechanical action also plays a notable part; in the same region underwater exploration has shown that boring organisms play an important part below low tide level, while they are also important in the intertidal zone of the coral seas. In hard and compact limestones, chemical corrosion is clearly the most effective of the processes acting on the part exposed at low water, and creates there a unique series of zones.

In non-calcareous rocks, there are often regular notches at the base

of some cliffs, which cannot definitely be attributed to mechanical erosion, e.g. in lavas and volcanic conglomerates of Fernando da Noronha, and the isle of S. Vicente in the Cape Verde Islands. As noted by Zahn (quoted by Panzer) notches are relatively rare in Brittany and the Isles of Scilly, that is, in places where they ought to be frequent if they were associated chiefly with mechanical processes. They are more regular and less deep than the hollows caused by breaking waves.

We must also point out, in addition to the forms already mentioned (p. 30), fine pittings in Lower Devonian shales in the Rade de Brest, in the Erquy shales of northern Brittany, and in the Ordovician and Cambrian shales of Morocco; pittings in the Permian rocks and porphyries of Provence (pl. I A); and small lapiés in aplite in the Rade de Brest. In general these features develop only in the spray zone; lower down the beach mechanical erosion predominates. Coastal lapiés in granite and granulite are fairly common, at least in Brittany on the coasts of Léon and Trégor, the island of Sein, Penmarc'h, and Concarneau, as well as in Hong Kong. They have the form of basins, channels, and gutters. Lines of weakness usually control their plan. But, even though they are essentially coastal and characteristic of the spray zone or a little farther back, they are also found in the interior of Brittany and Corsica (*tafoni*). Granular disintegration, possibly beginning in the Tertiary, may well have initiated such erosion. Sea-water and rain water may be the cause of their formation, but this is not yet entirely clear. There are also coastal lapiés in Cambrian conglomerates in south Morocco: these are due to differential erosion leaving the silicified parts of the rocks projecting.

Abrasion platforms caused mainly by mechanical erosion certainly exist in rocks other than hard limestone, notable in certain shales. Those with vertical or high dips at times have a polish comparable with that produced on marble.

But the detailed study of coastal platforms, especially by students of the Pacific coast, has revealed in Australia, New Zealand, and Hawaii features which strangely complicate the traditional picture of such forms. At numerous points on the coasts of these countries systems of minor rock terraces, which cannot always be explained by lithological influences, have been observed. The terraces are separated by very steep or even vertical slopes, veritable minor cliffs in the face of the major cliff. In the outskirts of Sydney, Jutson describes three platforms which he calls, from top to bottom, the high level, normal, and ultimate

platforms. The highest is only reached in general by spray and breakers; the second is in the tidal zone; the third is beneath low water level. Both in Hawaii and Australia the terraces are often bounded on the seaward side by a small, discontinuous, rock ridge, slightly higher than the terrace itself. These forms have been observed in volcanic tuff, altered basalts, ferruginous sandstone, arkoses, and shales: usually the rocks are decomposed or easily altered. Some Australian authors, especially Fairbridge and Teichert, liken these terraces to the succession of solution notches on limestone, and explain them by slight post-glacial eustatic changes of sea-level. But the 'Melbourne School' and Wentworth consid - this at most a partial explanation, and think that the terraces are all being formed at the present time; the sea, in effect, cuts not one but a series of erosion surfaces. Most investigators (Bartrum, Hills, Wentworth) consider that the highest flats are totally or partially due to 'water layer weathering' (see p. 30); mechanical abrasion would only be the dominant factor in the formation of the lower surfaces. The existence of the discontinuous outside ridge is explained by the fact that at that point the rock is more constantly saturated by breakers than it is elsewhere. If the platforms were due to mechanical abrasion, one would not be able to understand the preservation of this ridge in the most exposed part.

The efficacy of 'water layer weathering' is well established, although we have cast doubts (p. 30) on the exact nature of this erosion. But the part played by abrasion in the formation of raised platforms remains a matter for discussion. Edwards thinks it quite effective, even dominant, on those high water level platforms which are narrow. Hills says, moreover, that in Victoria the narrow high water level platforms are chiefly found where the waves are strong, the others in the sheltered parts. One might be tempted to conclude from this that decomposition at the edge of the pools, whatever its exact nature, is dominant, or at least very important, in the upper parts of the beach and the spray zone; lower down abrasion would play the chief part. The platforms at the level of high water would therefore be narrow where the lower abrasion platform cuts actively into them and wide where abrasion only reduces them slowly from the outside.

Nevertheless, there are two difficulties at least:

(*a*) If one admits that narrow high level platforms in exposed places are due to actual marine abrasion, one implies that the mechanical action of the waves is capable of creating a series of levels, a process very difficult to understand. The difficulty seems to be resolved only by

1A Corrosion pitting of Esterel porphyry at Anthéor in the spray zone. Note the effect of joints

1B Coastal lapies in Ordovician limestones at Lostmarc'h, Crozon peninsula, Brittany

1C Large flat-bottomed pools with residual lapies in Cretaceous limestone. Mazagan, Morocco

1D Large coastal overhang in coral limestone, Sherm Ubhur near Jidda, Red Sea. The hammer indicates the scale. (Photos A. Guilcher)

attributing *all* raised flats developed at the present day to some process other than mechanical abrasion.

(*b*) Waves are usually stronger at high water than at low water and exact measurements taken by Berthois have proved that on the coasts of Brittany, and probably elsewhere, these differences affect the rate at which loose stones and pebbles are rounded. Below mean sea-level on normal shores with concave profiles the process is extremely slow or non-existent, but is more rapid at the level of high spring tides and most effective near the level of neap tides and a little below: at the Pointe du Portzic (Goulet de Brest) the sea at this level has in less than six months transformed blocks of dumped sericite schists into rounded boulders comparable with the 'old' pebbles of the neighbouring shores. In such circumstances how could abrasion be preponderant at low tide level and be replaced higher up by weathering? This only seems conceivable in the case of a shore with a convex profile (p. 18) and it is not very likely that all the Pacific coasts which have been studied have such profiles.

These difficulties must not lead one to underestimate the importance of agents other than waves on the so-called abrasion platform. In addition to the development of solution notches, pitting and lapiés, the erosion on the edge of pools, the broadening of fissures and holes caused by marine organisms, we must also consider the action of ice in cold seas. There seems no doubt that the ledges, which surround many Norwegian fjords at about 50 cm. above mean sea-level, are due to coastal frost shattering. They cannot be explained by abrasion in such sheltered waters on coasts with so limited a fetch. Nansen does not hesitate to attribute to the action of the ice-foot the first stages in the formation of the strandflat (see p. 160). Without, therefore, going so far as to deny mechanical marine abrasion, we must associate with it or at times substitute for it other processes of mechanical, chemical, and biological nature.

The profile and recession of cliffs depend both upon the rocks in which they are cut and upon the forces acting on them, the wind, fetch, and depth of water in front, etc. Only the first factor will be considered here.

(*a*) In incoherent materials such as sand, e.g. eroded dunes, loess, glacial deposits, recent alluvium, and volcanic ash, evolution is generally swift and results from landslips (pl. II A and B). The cliffs of glacial drift in the region of Cromer and Mundesley in East Anglia offer examples of this. Sectors with a vertical or very steep profile alternate

with areas of hog back form resulting from the collapse of large masses. The speed of evolution is due to the fact that the deposits never attain a slope of equilibrium, because the sea is constantly clearing away any material which slips down. These slips which result from the tendency of the slope to become less steep, generally follow periods of heavy rain, as do mud-flows, and are, therefore, the result of subaerial rather than strictly marine action. But the removal of the debris from the base of the cliff by the sea is necessary for their continuation. The sectors with steep slopes are those where no recent slides have taken place. Cliffs of this kind can be seen in Brittany at Penestin, near the estuary of the Vilaine, in the loess of the coast of Trégorrois and in the region of St Cast, where the steepness is caused by the compactness of the loess. They are found in Denmark, e.g. the Bastrup cliff in Zealand on the Great Belt and are very common in New England, notably in the drumlins near Boston. In the drift cliffs of Norfolk and Yorkshire there are well known and often quoted examples of villages which have disappeared into the sea. Between Flamborough Head and Spurn Head at the mouth of the Humber, the drift coast of Holderness has receded on an average 65 m. between 1852 and 1889, that is about 1·75 m. a year. At Withernsea nearly 5·5 m. a year were lost between 1852 and 1876, and at another point on the same coast 305 m. were lost between 1847 and 1908 (Steers). But the record seems to be held by cliffs cut in volcanic ash: Umbgrove has shown that those of Krakatoa have in places receded 1,500 m. between 1883 and 1928. Similar facts have been noted by Hoffmeister and Ladd in the tufa, scoria, and pumice of the Tonga islands. Such erosion, together with subsidence, is the explanation of the rapid disappearance of small volcanic islands like the one seen near Pantellaria in 1831.

(*b*) In clay, the cliff may be cut by ravines of the type found in bad lands (Fig. 4 I): the lower part is formed, at least in places, of convex masses which spread over the beach during periods of heavy rain. The clay absorbs the rain (Fig. 4 A), and then dries out and becomes riddled with cracks due to contraction. Here again, the form is of subaerial origin, but the continuation of the flows depends on the washing action of the sea at the base. Sometimes on the reverse slopes between the slipped masses and the cliff proper pools are formed. The cliff of the Vaches Noires in the Oxford Clay in Normandy shows typical bad lands, as does that at Boulogne in the Kimmeridge Clay and that of Highland Light near Cape Cod in New England. The convex masses and the hollows from which they have slipped are very clear in the

plastic clay at Rjosnaes in the west of Zealand in Denmark, and at Trimingham, Norfolk, in the marly and compact till at the base of the Pleistocene (these beds are attributed to the North Sea glaciation and lie below more sandy layers). Clay cliffs recede generally much less rapidly than the preceding type, but at Trimingham the top of the cliff, being more sandy, undergoes slipping. There is here, therefore, a superposition of the two types.

(*c*) A very interesting type of cliff occurs where a hard massive rock lies above an incompetent impermeable rock. Such an arrangement is seen at the Warren at Folkestone, at La Hève near Le Havre, and at Seaton in Devon (Fig. 4 B and pl. II C). In the first and second places, the upper part is Chalk, and the base, at least in part, clays of Middle and Lower Cretaceous or Upper Jurassic age; in the third place the Chalk and the Upper Greensand form the compact higher part, while below are sands and clays, ranging from Lias to Keuper in age. Such cliffs are subject to huge landslips; those of Christmas 1839 in the neighbourhood of Seaton involved a mass estimated at 8 million tons; in the Warren at Folkestone, ten important slides occurred between 1765 and 1915. The chalk at the top exerts a pressure on the underlying clay: when the clay is saturated by prolonged rains, it flows towards the sea, carrying with it huge blocks of chalk. At the Warren very low tides create a situation favourable to slides since the base of the cliff is not then supported by the sea. The profile is, therefore, one of a staircase of faulted blocks, called Panamanian faults by J. Bourcart, with more or less convex masses at the base. The details of the slides vary with the locality and peculiarities of structure; but very large falls in chalk cliffs are strictly confined to places where the base of the cliff is not chalk but clay.

When, on the other hand, saturated clay is found above hard rocks, the cliff has a convex outline, the mud flowing over the edges of the hard beds which may at times be concealed, e.g. at Folkestone between the town and the Warren, where the base is formed of Lower Greensand and the top of Gault Clay.

(*d*) Profiles and modes of evolution in resistant rocks are not everywhere the same. In every case retreat is less rapid than in the preceding types, but it varies from rock to rock.

Calcareous rocks including chalk tend to give vertical cliffs, as do columnar basalts, sandstones and certain old shales (Fig. 4 C and F). Examples are found in the Pays de Caux and in south-east England in chalk, in the north of Scotland at Duncansby Head and in the Orkneys

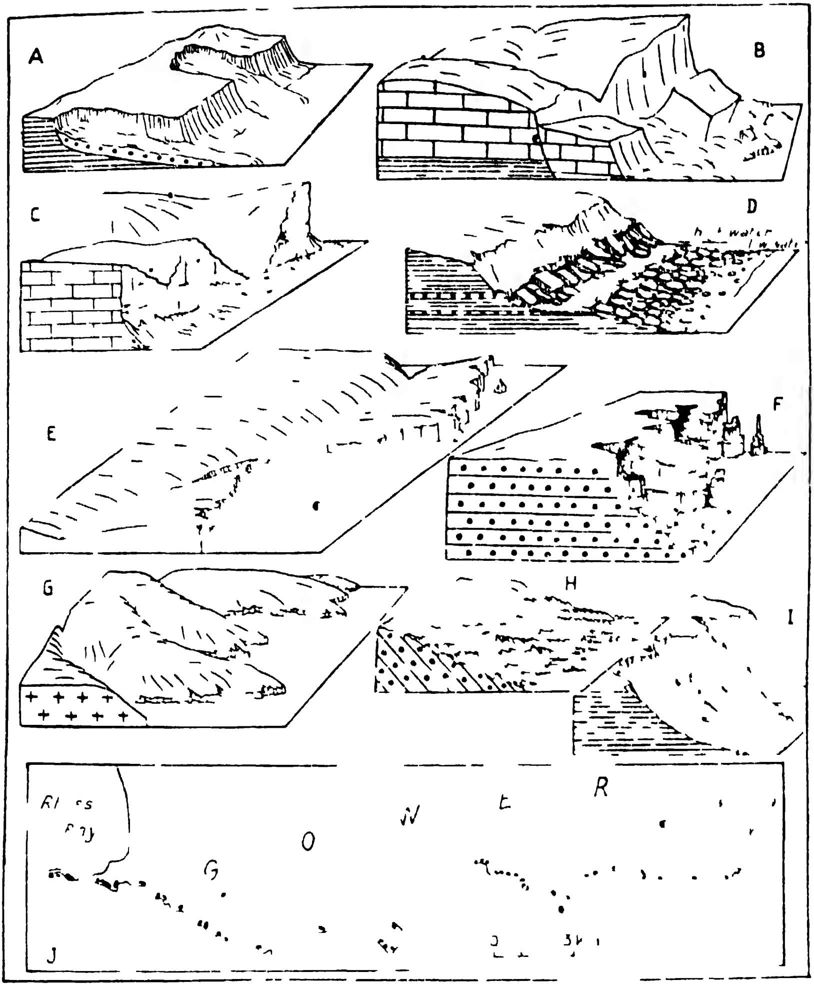


FIG 4 CLIFF TYPES

A Clay cliffs with mud flows (after Schou) Rjosnaes, Denmark, type. B. Cliffs with large land slips limestones over marls Seaton type C Chalk cliff with talus and a dry hanging valley Caux type D Cliff in interbedded flagstones and clays, with slabs of flagstones resting on the clay Alprech, Boulonnais, type E Valley captured by cliff recession and so rejuvenated F Cliff characteristic of horizontal sandstone, limestone or basalt Frehel or Duncansby type G False cliffs with small modern sea-cliff at base ancient massif, or Côte d'Azur type H Cliff in sandstones, limestones, or slates with strong seaward dip Frehel type I. Bad lands and waste fans developed on clay cliffs J Permanence of Carboniferous Limestone cliffs of the Gower peninsula indicated by the presence of the raised beach (black dots) at their foot (after George, 1932).

in Old Red Sandstone, in Ireland in basalts, in Brittany at Belle-Ile in pre-Cambrian schists, and at Cap Fréhel in Carboniferous sandstone. In every case there are exceptions to the rule, particularly when the dip is towards the sea and is not too near the vertical; either the cliff profile follows the dip, or, when the beds dip with, but at a greater angle than the slope, the profile takes on a saw-tooth appearance, for example between Cap Fréhel and Sables-d'Or-les-Pins (Fig. 4 H). Horizontal beds, beds dipping landwards, or almost vertical beds seem to offer conditions favourable to vertical cliffs: but structural details and local differences in evolution often lessen this verticality. The steepness of chalk cliffs, for example, may be decreased by landslides, as between Dover and Folkestone. All these cliffs evolve as a result of violent and infrequent slides, while in the intervals between little or nothing happens. The fallen masses enclose great boulders, which come from high up in the cliffs and have been for a long time subjected to erosion by land water in the joints and cracks. In general it may be said that all the preceding types of cliffs only evolve visibly at times of storms accompanied by large waves and heavy rain, which occur mainly in autumn and winter in oceanic temperate countries. Such discontinuity of evolution is characteristic of mechanical action on the coast, and is the opposite of what happens with chemical erosion.

Stacks and arches in front of chalk cliffs are somewhat exceptional. The Sheringham tubular stacks are associated with chalk hardened by secondary lithification along solution planes (Burnaby).

In crystalline and metamorphic rocks, and in very hard sandstone like the Armorican sandstone of Ordovician age, the 'cliffs' have a convex outline and are usually covered with vegetation over the greater part of their height. Only the lowest third or quarter is subject to the direct action of the waves and forms a true cliff; the rest is a slope of entirely subaerial origin and must be called a false cliff (Fig. 4 G). It is the typical cliff profile of ancient massifs such as Wales, Cornwall, the Armorican Massif, the Maures Massif, Cintra, and Nova Scotia. In the part attacked by the sea, recent erosion may be pronounced, but it remains a detail only. The profile suggests stability: it is really the slope of a hill eaten into by the sea at its base without any modification being caused higher up. This is equally true in Provence, along the coast between Marseilles and Cassis, which is formed of hard Lower Cretaceous limestone. Even the high escarpments between Cassis and la Ciotat are not true cliffs but a cuesta enclosing the basin of le Beausset: scarps just as pronounced occur in certain places not attacked by the

sea north of Cassis.[1] False cliffs again occur on the high coasts of Cap Roux in Estérel, Cap Ferrat in Villefranche, and Cap Martin near Menton (pl. II D). Old massifs are not, therefore, the only places where the mechanical action of the sea seems weak today.

In the Armorican massif the platform in front of the false cliff is not a true abrasion platform: it is littered with angular blocks produced by the waves re-sorting periglacial Quaternary solifluxion material resting on solid rock, which is itself heavily fissured by cryoturbation contemporaneous with the flows. For example, at Plouguerneau and Guisséný in north Finistère, it is easy to see the transition from intact periglacial deposits to the residual deposits of large boulders which litter the shore.

A consideration of the relief of the land in the immediate vicinity often supplies valuable indications of the importance of marine erosion. If the land is hilly and if a true cliff is backed by land sloping away from the sea, the sea has consumed at least half the hill. Where the slope above the cliff is directed towards the sea marine erosion has not yet destroyed half the hill, and by extrapolating that slope one can make an estimate of the amount of land lost, unless one or more hills lying in front of the cliff have already been totally swept away. Generally, however, an examination of a fairly long sector of the coast by an experienced observer will allow him to decide what has happened.

In regions of boulder clay and drumlins a complete disorganization of the relief of the land is caused by the retreat of the cliffs: the drumlins are in all stages of erosion; numerous examples occur in New England. In chalk it is not rare to see hills half eroded away; the valleys cannot adapt themselves to the rate of cliff recession and end in abrupt drops, thus forming hanging valleys; but generally a good aerial photograph allows the general position of the shore previous to the retreat of the cliffs to be reconstructed by prolonging the slopes of the hills. In old massifs, either there is no retreat or only negligible retreat: valleys parallel to the shore and captured by small marine inlets (Fig. 4 E) are not uncommon on some coasts but they are by no means typical. They occur at Belle-Ile, Ushant, south of Kingsbridge near Start Point, and especially in north Devon and north-west Somerset where Arber has

[1] These are not the highest cliffs of France, as has sometimes been claimed. The highest cliffs are found at Houlgate in the pays d'Auge in Oxford Clay and Lower Chalk; but particularly in Caux between la Hève and Antifer, just north-east of Fécamp, between Dieppe and Criel, and at Blanc-Nez in the Pas-de-Calais—all these are chalk. All are roughly 110–15 m. high.

made an excellent study of them: the Valley of the Rocks at Lynton is a model of this kind.

Another proof of the slowness of retreat in many places is the presence at their base of strips of beach which existed before the last glaciation, the so-called Lower Monastirian beach. These beaches are numerous all over western Europe (pp. 45-6). They show that many shores, notably the north of the bay of Audierne, in the Gower peninsula of Wales, and in south-east Ireland, have receded little since the end of the Ice Age (Fig. 4 J). The beaches rest on an old marine platform, which was probably never very wide. Such beaches are not unknown outside the ancient massifs: they can be found in the Chalk on the outskirts of Brighton, on the east coast of the Straits of Dover, and in Picardy. In many other parts of the world, e.g. Morocco, Algeria, Portugal, and the Lebanon, the existence of old raised beaches along the coast shows the feebleness of marine erosion at present sea-level. Rocky shores are often old shores again subjected to erosion. It can, therefore, be said that the rapid retreat shown by cliffs of glacial drift and volcanic ash is rather exceptional. A large number of authors have reached the conclusion that erosion of rocky coasts is much slower than subaerial erosion; Wentworth (1928) considers it to be seven times slower in the Hawaiian Islands; Smit Sibinga, from a consideration of the relative rarity of hanging valleys, is equally favourable to the idea of the dominance of fluvial denudation; if this were not so, he says, there would be coastal falls along all the coasts in the world.[1] Kuenen reaches similar conclusions.

We must not, however, be dogmatic in this matter. On the west and south-west coasts of New Zealand, Cotton points out rapid retreats in many places. Although these are not measured, they are shown quite clearly in photographs, and occur even in resistant rocks like those of the volcanic district of the Waitakere range near Auckland. We must, to understand them, take into account the great swell of the west and south Pacific, which, because of the steepness of the submarine profile, is hardly weakened before reaching the New Zealand coast.

When the sea leaves a shore as a result of a movement of base-level or because of aggradation at the foot of the cliff, the cliff rapidly becomes degraded. There are certainly old marine notches on many coasts, while dead cliffs like those of the Bas-Champs of Picardy are not rare. But

[1] Certain hanging valleys are not associated with the retreat of the cliff, but result from changes of sea-level. Examples are found in south Brittany, in some Cornish valleys and in the Cintra Massif in Portugal (Guilcher, 1948 and 1949).

their forms are degenerate, their slope diminishes quite quickly and those which are older than the last glaciation have often been covered with flows of periglacial solifluxion or with loess. Where they are now fresh-looking, they have been re-attacked by the waves.

B. BEACHES AND COASTAL DUNES

'The sea is a consumer and not a producer of sand.' Ferronnière's striking phrase illustrates the great importance of constructional features. Unlike most cliffs, beaches are built, altered, and eroded with great rapidity. In beaches much more than on cliffs there is continual change.

The materials of beaches. A beach can be defined as an accumulation on the sea-shore of material coarser than mud (pp. 100 1). But certain beaches become muddy in the low water zone. The following terminology has been proposed by Bourcart for the calibre or grades of material:

—from 1 m. or more to 2 mm.: *pebbles.* This category includes *boulders* (diameter above 500 mm.), *shingle and angular pebbles* (500 to 25 mm.), *gravel* (25 to 10 mm.), and *fine gravel* (5 to 2 mm.).

—from 2 to 0·02 mm.: *sand.* This includes *coarse sand* (2 to 1 mm.), *medium sand* (1 to 0·5 mm.), *fine sand* (0·5 to 0·1 mm.), *very fine sand* (0·1 to 0·02 mm.).

—from 0·01 to 0·001 mm. (*silt*): from 0 001 mm. to 0·0001 mm. (*pre-colloids*): less than 0·0001 mm. (*colloids*).

A natural sediment nearly always includes several grades. Beaches range in texture from a predominance of the coarsest material to a predominance of very fine sand. The term shingle beach should be restricted to one with a predominance of the coarse grades, shingle and gravel.

There is no progressive change from one grade to another by gradual attrition: in particular, sand appears to be derived only to a small extent, if at all, from the reduction of shingle. By collisions shingle breaks up to produce smaller shingle; by rubbing shingle produces silts or colloids. Sand may be eroded by the sea from river deposits; from rocks already rotted, from uncemented sandstones, or from glacial or periglacial deposits. But only the sea is capable of quickly giving them that smooth roundness which has been defined by A. Cailleux; the sea

IIA Cromer cliffs in glacial deposits with chalk erratic 12 m long and landslips

IIB Drumlin cut in two by the sea, Youghal Bay, near Cork

IIC Coastal landslips at the Warren, Folkestone.

IID False cliffs at Cap Martin, Alpes-Maritimes
(*Photos A. Guilcher.*)

produces in the stones a higher index of roundness, a greater flatness and greater symmetry than do rivers.[1] By these characteristics marine sands and pebbles can be distinguished from fluvial deposits.

The contrasts in the rapidity of the rounding of stones at different heights on the beach, as found by Berthois, have been emphasized above (p. 69). This depends on the concavity of the beach which reduces the breaking of the waves at low water and also on the *Laminaria* which grow at very low tide in middle and high latitudes (p. 31). The exceptions which occur when the profile is not concave concern rock platforms rather than beaches.

In low parts, however, certain organisms which pass material through their intestines can transform calcareous sands into mud (p. 31). The sea may also continually sweep on to certain coasts, like those of the Landes of Gascony, masses of sand the source of which is not always known. Heavy mineral analysis, however, is often useful in establishing the origin of the material as Vatan did on the coast of the Gulf of Lions.

The materials of a beach are hardly ever even approximately of the same size: they are heterogeneous, but better sorted than fluvial sediments. The sorting of the material on the basis of density, size, and form, and the fragmentation of shells, is more complete where the beach is very exposed to the sea. When the deposits includes shingle and sand, the shingle is mainly concentrated at the surface in the zone of high tide and storm waves, in the low tide zone sand alone or sand with little else is present; in the same zone and in that of very low tide angular pebbles are mixed with the sand on certain beaches or during certain periods. In vertical sections through the beach in the shingle zone, the shingle below is smaller than on the surface, and mixed with gravel. Sections in sand and gravel without shingle are more complex. Certain rocks favour the presence of banks of shingle, e.g. chalk because of its flint, and all sedimentary rocks that resist destruction. As a result there are large numbers of flint shingle beaches on the coast of the Pays de Caux and south-eastern England, shale or slate beaches in Cardigan Bay, and beaches formed from Armorican sandstone on the Crozon peninsula. On the coasts of granite massifs which have been largely weathered into sand, shingle beaches are, on the contrary, rare,

[1] Definitions (Fig. 5 B): L = the length of a stone, l = its breadth, E = its thickness, r = the least radius of curvature in the principal plane of the stone. The asymmetry is $\frac{AC}{L}$; the flatness $\frac{L+l}{2E}$, the index of roundness is $\frac{2r}{L}$.

e.g. north Brittany, and sand usually dominates up to the top of the beach. But in certain cases accumulations of shingle may be derived from considerable distances, perhaps brought by floating ice in the Quaternary period.

Stones scattered over a sandy beach have their main axes parallel to the shore, that is parallel to the waves (Cailleux). Currents and waves on the coast cause the stones to overlap in two different ways (Fig. 5 C). This may reveal, in coastal accumulations of shingle where currents and waves operate at the same time, the predominant cause of the arrangement. It has been observed that waves are the more important, except in certain cases where a tidal current always flows in the same direction, as at Kilaourou, island of Sein.

On the banks of certain lakes winter frost-shattering prevents the formation of shingle as there is insufficient time for rounding to occur. Taber has established this in the United States in the state of New York, but only where stones are embedded in clay: those embedded in gravel are rarely cracked and are formed into shingle.

The essential parts of a beach. We may propose the following classifications of the different parts of a beach (Fig. 5 A):

—The offshore zone (Fr. *avant-côte*): situated in front of the beach.

—The level of the highest tide in calm water (Fr. *trait de côte*); a feature which on maps marks the limit of land and sea.

—Upper beach (Fr. *cordon littoral*) and lower beach or foreshore (Fr. *bas de plage*): the two parts of the beach (Nos. 3 and 4) are often separated by a line (No. 5) where the slope and calibre of the deposits decrease markedly. When the upper beach is not tied to a cliff or dune line it includes beyond the beach a reverse slope (No. 6) and in this case often encloses a lagoon (Fr. *étang de barrage*).

—Beach ridges or berms (Fr. *gradins de plage*): ridges built up by successive high water levels, or, in tideless seas, by storms of decreasing height (No. 9). These are particularly clear on shingle beaches, where up to 5 or 6 ridges may be recognized when the beach is very high, e.g. Ero Vili in the Bay of Audierne, Chesil beach near Portland, and Dungeness. The highest ridges (Fr. *crête de plage*) are built up by large storm waves at some height above the high water level of calm water (No. 10). The crest of Chesil beach reaches 13 m. above high water level in its south-east portion. This is exceptionally high and is due to its

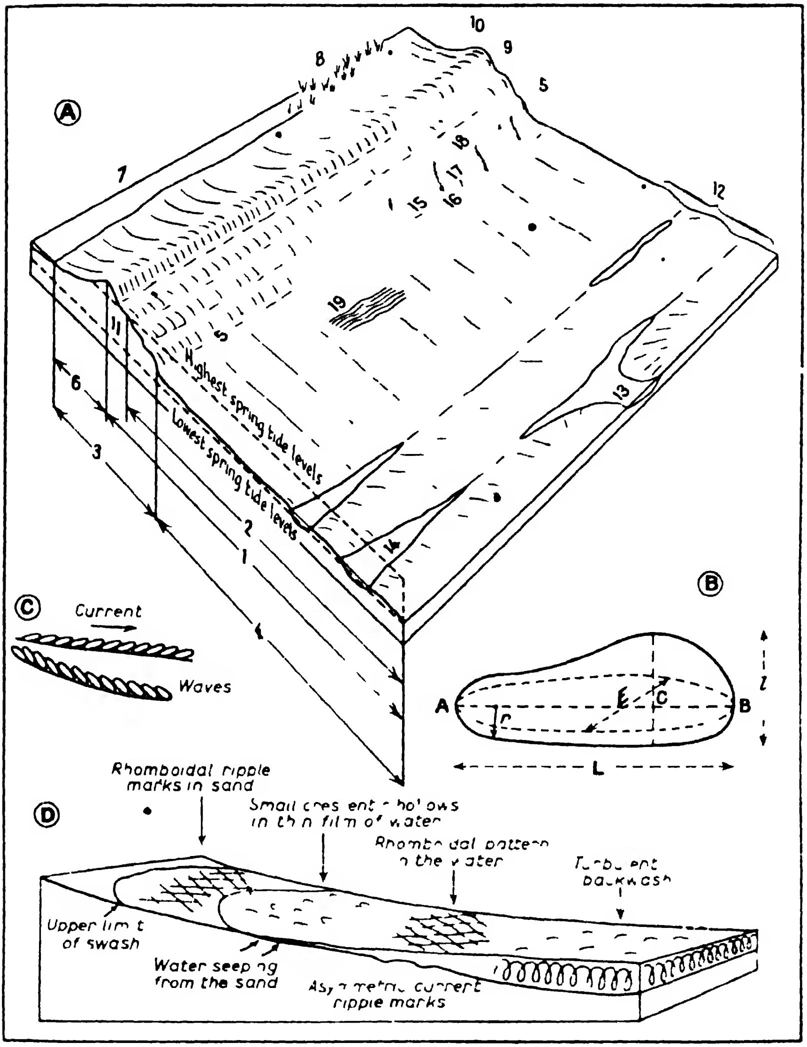


FIG. 5 BEACH TERMINOLOGY

A. 1. Beach. 2. Shore. 3. Upper beach (cordon littoral). 4. Foreshore. 5. Break of slope between upper beach and foreshore. 6. Inner side of beach ridge. 7. Lagoon. 8. Marsh. 9. Berms. 10. Storm beach. 11. Coastline. 12. Ridges and runnels on the foreshore. 13. Channel on foreshore. 14. Pool in runnel on foreshore. 15. Beach cusp. 16. Apex of cusp. 17. Bay of cusp. 18. Horn of cusp. 19. Ripple marks.

B. Dimensions of a pebble: L = length; l = width; E = thickness; r = least radius of curvature in the principal plane (after Cailleux, Berthois, and Tricart).

C. Ways in which pebbles may overlap under the influence of waves and currents (after Twenhofel).

D. Formation of rhomboidal ripple marks (after Demarest).

exposure towards the south-west facing the direction of greatest fetch (Fig. 27 A).

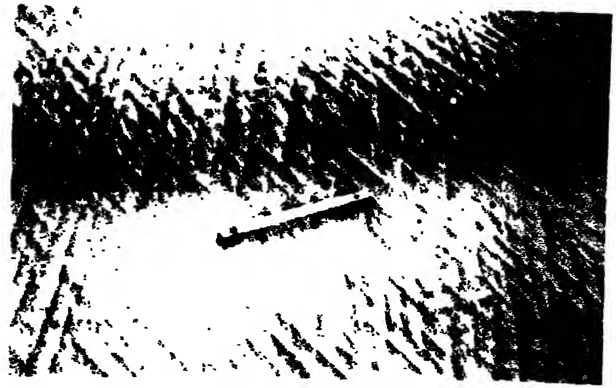
The slope of a beach varies with (*a*) the material of which it is composed, being steeper on shingle than on sand, (*b*) the part of the profile under consideration, as gradients are steeper in the upper parts of the beach, (*c*) the local topography, gradients being steeper near rocks. Shepard quotes mean gradients ranging from 2 in sand, 0·12 mm. in diameter, to 20 in pebbles of 64 mm. diameter. We have measured in Morocco exceptional slopes of from 30 to 31 on certain shingle beaches and 11 on certain sand beaches, where no undercutting has occurred; the first of these figures agrees with that given by Kuenen. Bascom, who has studied the profiles of forty sand beaches, with an average grain size of 0·17 mm. to 1 mm., on the Pacific coast of the United States has found, in the zone subjected to the action of the waves at half-tide in a number of places with different exposures, slopes of from 1 to 22 . In the same zone erosion of the beach flattens the profiles, aggradation steepens it.

Minor beach forms

Beach cusps (*Fr. Croissants de plage*). These are features of the upper beach and not the foreshore (Fig. 5 A, Nos. 15–18). They are found in tidal seas, tideless seas, and lakes. They consist of triangular shingle or gravel ridges separated by hollows open towards the sea, and occur in series of variable length. The difference in height between the hollows (bays) and the crests (horns) varies usually from a few centimetres to a few decimetres; occasionally it exceeds 1 m. and we have seen one case at St Ives (Cornwall) of a difference of 3·40 m. The wave-length (distance between two crests or two consecutive hollows) is also variable; the average wave-length of twenty-two series measured by the author varied from 5 m. to 77 m. In any one series there are sometimes considerable irregularities. An important point is the effect of the material on the size and regularity of the cusps: other things being equal the cusps are longer and more regular in sand than in shingle, and longer in shingle than in sea-weed when this is entangled in the upper beach. The reason is, doubtless, that sand is easier to move than shingle. Yet cusps are rarer in sand than in shingle, perhaps because a fairly strong slope is necessary for their formation, as Escher has affirmed, and because shingle beaches have usually steeper slopes than sand beaches. But, in a given set of cusps, the material is coarser in the

IIIA Storm beaches at Pwll Du, Gower peninsula, South Wales

IIIB Two sets of ripple marks superimposed on the beach, near Fedala, Morocco



IIIC Rhomboidal ripple marks at Montalivet, Landes, Gascony. (*Photos. A. Guilcher.*)

horns than in the bays, as Guilcher, Berthois, and Boyé pointed out. Boyé attributed this to the process of their formation.

Cusps are associated with the movement of the swash. All systems of large waves may not produce large cusps but all large cusps are formed by large waves; when there are several systems built up successively on a beach at different heights, the highest are generally the largest as they are due to the greatest waves. The breaking wave rushes up into the hollow: during the backwash, it retreats first from the cusps and then from the hollows. The backwash of a large wave acts as a brake on the following smaller wave which advances less into the hollows than towards the cusps.

The process of formation is not yet well known. Several authors (Jefferson, Evans, Guilcher) have asserted that beach cusps may be formed from the breaking up of banks of sea-weed or from a beach ridge; this break localizes the outflow from the spray-fed pool behind the ridge. But the breaches must be spaced at intervals corresponding to the wave-length of the cusp system which can establish itself at the place under consideration as a function of wave-height, the slope of the beach, and the material of which it is formed. This is doubtless not the only method of formation. On the other hand, it has not yet been explained why, with similar slopes and material, certain beaches have cusps and others have not. We can only observe that certain beaches develop them more readily than others, and that they keep them more or less permanently, e.g. the pebble beach in the Bay of Audierne. Rivière and Mademoiselle Vernhet have noted that cusps form as the sea calms down after a storm in Provence; but that does not appear to happen everywhere. According to Boyé, they are formed at the beginning of the ebb-tide in seas where the tidal range is large.

Ripple marks (*Fr. rides de plage*) are found on the foreshore and not the upper beach (Fig. 5 A, No. 19). Sand ripples are found at the bottom of lakes, rivers, and seas at very great depths, even as low as 1,500 to 2,000 m. in restricted straits in the Canaries and East Indies, according to Twenhofel.

Sand ripples are of different types. Oscillatory ripples have symmetrical slopes, sharp crests, and rounded troughs. They result from the to-and-fro movement on the sea bottom produced by the passage of waves. Their size varies with that of the waves and the depth of water. Their average height is from 5 to 15 mm.; their wave-length 3 to 12 cm. They are perpendicular to the direction of the movement of the waves.

Ripple marks produced by currents (pl. III B) are asymmetrical and their crests are rounded. They are formed at right angles to the current, and their gentle slope faces the direction from which the current comes, the other slope being the angle of rest of the material. They are formed when the current exceeds a critical speed, varying with the depth of water and the size of the sand grains, and they move in the direction of the current. If the current exceeds a second critical point, they cease to be formed, as the sand is lifted and moves in suspension in the water instead of rolling on the bottom. If the speed increases further and reaches a third critical point ripples are again formed, but are displaced upstream by erosion on their downstream slopes and deposition on their upstream slope; these ripples, which are symmetrical, have been called 'antidunes' by Gilbert. The passing of the third critical point leads to the formation of broad, rounded, and more irregular crests which move downstream with the current. A decrease of speed changes them into current ripple marks of the first kind. Ripple marks at great depths are all due to currents. This type is common in rivers and estuaries; the crests of those of the estuarine part of the Loire, called '*ridens*', move a few metres both downstream and upstream during one tide. The play of swash on a beach can also create transitory current ripple marks between two waves. Those of the Loire can exceed 4 m. in height; those on the sea-floor, which are never seen, may reach 8 m. or even 13 m., as on the Falls and the Varne banks, in the southern North Sea and Straits of Dover.

Lozenge-shaped or rhomboidal ripple marks (pl. III C) quite often form systems of marked regularity. The rhombs, the major diameters of which follow the slope of the beach, are only a few centimetres long. They are formed at the time of the backwash, as described by Demarest (Fig. 5 D). The flow of the film of water in rapid retreat is at first very turbulent. When it does not exceed approximately 2 cm. in depth, a rhomboidal system is formed in it; when the depth is no more than 1 cm. small current ripple marks are formed in the sand; then everything becomes indistinct. After the backwash has passed, water springs from the saturated sand; small temporary cusps, like little ditches concave towards the lower part of the beach, are formed in the sand. Then, when the sand dries, a rhomboidal system appears. All this happens in a matter of seconds.

All the preceding ripple marks are forms affecting sand or gravel; but Van Straaten has also mentioned the existence of ripple marks in mud on the shore of the Wadden Sea in Friesland. They are sym-

metrical and parallel to the direction of flow of the tidal current unlike the ripple marks of the neighbouring sand banks which are perpendicular to the current and asymmetrical. They disappear in very calm weather.

It is common to see oscillation and current ripple marks combined as well as two sets of the same type crossing each other. These compound forms are more common than simple systems. When the two causes are not of equal importance the secondary system is superimposed on the principal one, which dominates the pattern with its greater wave-length and height. The complexity of sand ripple marks is therefore considerable; it is, in fact, greater than can be dealt with here. Although generally temporary, these formations may be preserved in fossil form; very fine fossilized ripple marks are seen on the Trias flagstones of the dome of Barrot in the Maritime Alps, and on the Ordovician of Dar bou Azza near Casablanca.

Lows and balls, runnels and ridges (*Fr. crêtes et sillons prélittoraux*). These forms are rarely met with on the upper beach but more often on the lower part of the beach (Fig. 5 A, Nos. 12–14) and offshore in seas and lakes. They are ridges parallel to the general direction of the beach or slightly oblique in relation to it; their slopes are gentler than those of the ridges of the upper beach. However, the difference of level between crest and hollow may reach or exceed 1 m. and even several metres in forms which are always submerged. The continuity of those which emerge at low water is often interrupted by small channels. In the runnels ponds may be found, from which water runs away through the channels at low tide. Lows and balls are very common on broad beaches of sand, e.g. the coast of the Landes, the Picardy Bas-Champs, southern part of La Vendée, French and Belgian Flanders, Blackpool and Southport, Blakeney (Norfolk), Norderney (East Friesland), the Algerian coasts notably at Sidi Ferruch, and Lake Michigan. Some are relatively fixed in position, others tend to move towards the upper beach and to disappear there (migrating bars), a fact which has been observed in tidal seas by Doeglas in Holland and even in tideless seas by Schou on the Baltic. All vary in height according to the weather. Generally speaking these variations are linked with the strength of the swell. In severe weather the crests, whether they are parallel or oblique, flatten or disappear; they form again in a moderate swell.

Williams considers that one must distinguish, from the genetic point of view, bars in tideless seas from those in tidal seas. The first—and this theory is supported by Evans—are formed at the break-point of waves; the wave makes a small hollow, the sand removed being formed into a

bar on the seaward side. Timmerman's experiments in a wave tank seem to have confirmed this theory. Systems of waves of different heights not breaking at the same point might explain compound systems.

In tidal waters it is clear that this explanation cannot carry much weight as the break-point is constantly shifting with the tide. Williams thinks that the cause might be the swash, which at the time of the ebb would build up a series of small ridges enclosing pools: these ridges would build up again at each ebbtide. This explanation does not appear very plausible; in any case it takes no account of the constantly submerged ridges which form part of the same series. It would be awkward to have to assign a different origin to these. The problem has not been solved, nor has the absence of these forms on, and in front of certainly apparently favourable beaches been satisfactorily explained. It is not correct to say that they exist only on beaches with a very gentle slope, because they are met with on the coast of the Landes. The channels which interrupt the ridges on the lower beach may be due in places to rip currents (pp. 20 1) and so may the parts of the troughs adjacent to the channels. This has been seen at La Jolla in California.

Offshore bars parallel to the coast do not appear to be formed of material brought by longshore drift. This idea (Gilbert, 1885, pp. 88–89) has been rightly opposed by Johnson, Timmermans, and Doeglas.

Ridges oblique to the shoreline appear to be associated with the predominance of an oblique swell: e.g. north-westerly swell on the coasts of the Landes, Senegal, and south Mauretania, and south-westerly swell on the coast of the Bas-Champs of Picardy. The ridges tend to align themselves perpendicularly to the swell, thus illustrating a general tendency in coastal forms (Lewis's law: p. 182). In these examples, a longshore movement of material by coastal drift is common.

The changing profile of equilibrium of beaches. It is a well-established fact (H. Baulig, Mlle Lefèvre) that the profile of equilibrium of a river is not necessarily a final profile which suffers no further change unless base level changes. This is even more true of the profile of equilibrium of beaches, where a balance is attained between the dimensions of the waves, and the calibre and amount of the material derived from the sea-floor, cliffs, or rivers. In general the profile is concave but in detail it may very well diverge from a regular concavity, especially when affected by ridges and runnels. The profile of equilibrium of beaches is attained much more quickly than that of streams, and is constantly modified, especially by variations in the strength of the waves. It does not depend upon the youth or maturity of the shore; if the latter con-

tains no outcrop of solid rock, a profile of equilibrium will be formed no matter what stage of evolution the shoreline has reached.

In particular, beaches are combed down when storms blow onshore, and built up in times of fine weather or offshore winds. To put it another way the profile is lowered when the waves are large and rises when they are small. Variations in the amplitude of the ridges and runnels (p. 83) are associated with this phenomenon: examples are as numerous as the beaches themselves. For example, during the 1939–45 war an abnormally long period of north-east winds lowered the shingle ridge at the edge of the breakwater south of Fécamp by 7 m., so that the foundations were laid bare and the German occupation authorities had to repair it. But scarcely had the repair begun than the prevailing south-west offshore winds returned and the beach aggraded with great rapidity so that protection became unnecessary. Combing down can remove the whole of a beach and cause the underlying rocks to appear. This happened at Deauville in 1950, when the mud underlying the sand appeared at the surface. Material carried away by storms never appears to go very far; it is transferred offshore for a time, or even only to the lower parts of the beach, and then comes back again. At Safi, for example, heavy north-west swell, which erodes the beach, deposits material on the sea-floor in front down to a depth of 7 m. below sea-level; lower down, variations are small. In the United States the *Beach Erosion Board* has made a systematic study of this type of erosion in order to try to find a remedy for it. These studies have considerable economic value, e.g. to the tourist industry. Some forms of erosion continue for several years as on the Normandy beaches after 1950; their causes are still not understood.

Major shore features. More or less permanent shore features are numerous. The most common is the bay-head beach which is arcuate unless there are rocky outcrops in the bay: these may cause the beach to form a series of arcs between the rocks. The form of bay-head beaches reflects the waves which shape them: refraction causes the waves to break in arcuate form, for the submarine contours are usually nearly parallel to the contours of the land and the waves which approach the bay normally tend to conform to these submarine contours. Regularity is attained by constant reworking of the bottom sediments near the shore, and the arc, or series of arcs, becomes perfectly geometric. Beaches occupying very large bays like that extending from the Pointe de Grave to Bayonne are not genetically different from the others. The orientation these beaches tend to take will be examined in Chapter V.

Behind the bay-head beaches there may be a cliff which the sea attacks at high tide or during storms and which retreats in the usual way. It would be a great mistake to think that there is nowhere any permanent erosion in bays (see p. 189). The beach, supplied with detritus worn from the cliff, is progressively retreating landwards, as for example in the bay of Audierne between Audierne and Pouldreuzic, in the bay of Aberdaron in north-west Wales and generally in small inlets in ancient massifs.

When, on the other hand, the beach is prograding, several successive beach ridges may be left. A very fine example is seen at Pwll Du on the Gower peninsula near Swansea (pl. III A) where the sea has built five ridges, which it has already abandoned, and is now in process of constructing a sixth. The sea thus tends to fill up the bay. Much bigger examples are found on the coasts of Gabon, of Portuguese and French Guinea, and the east of Madagascar: these are called coasts of 'parallel formations' by P. Legoux.

If a stream flows into the head of the bay and if it is not very powerful, its mouth may be barred by a beach and a pool forms. A narrow channel may be maintained, as at Arcachon, or the pool may be completely barred like the *loc'hiou* (singular: *loc'h*) at the south of the bay of Audierne. These barred pools are often emptied only by percolation through the beach, as at Pwll Du and south-east of Concarneau. This is easy when the barrier is formed of shingle. When the river rises, the pool may overflow and scour out a deep bed, but the sea generally closes the breach afterwards. Cycles of opening and closing may be established, and are sometimes aided by man, as in the bay of Audierne. Barred pools often flow into one another, and may have only one outlet: such are those of the Landes, where the outlets are called *boucaus* (Fig. 6 D), and the one south-east of Lorient which flows into the Gavre lagoon.

A spit of sand or pebbles perpendicular to the axis of the bay may occasionally be found half-way between the head and the mouth of a bay: this is Johnson's midbay bar, several examples of which he has quoted from the United States. The cove of l'Auberlac'h in the Rade de Brest, and the one at Croisty in the Rhuys peninsula (Fig. 6 A) afford typical examples of the form. According to Zenkovitch, the first stage in their formation in Kamchatka is due to the progressive retardation of waves moving from the open sea into bays, thus leading to the deposition at a certain point of the coarsest fraction of the material they bring with them.

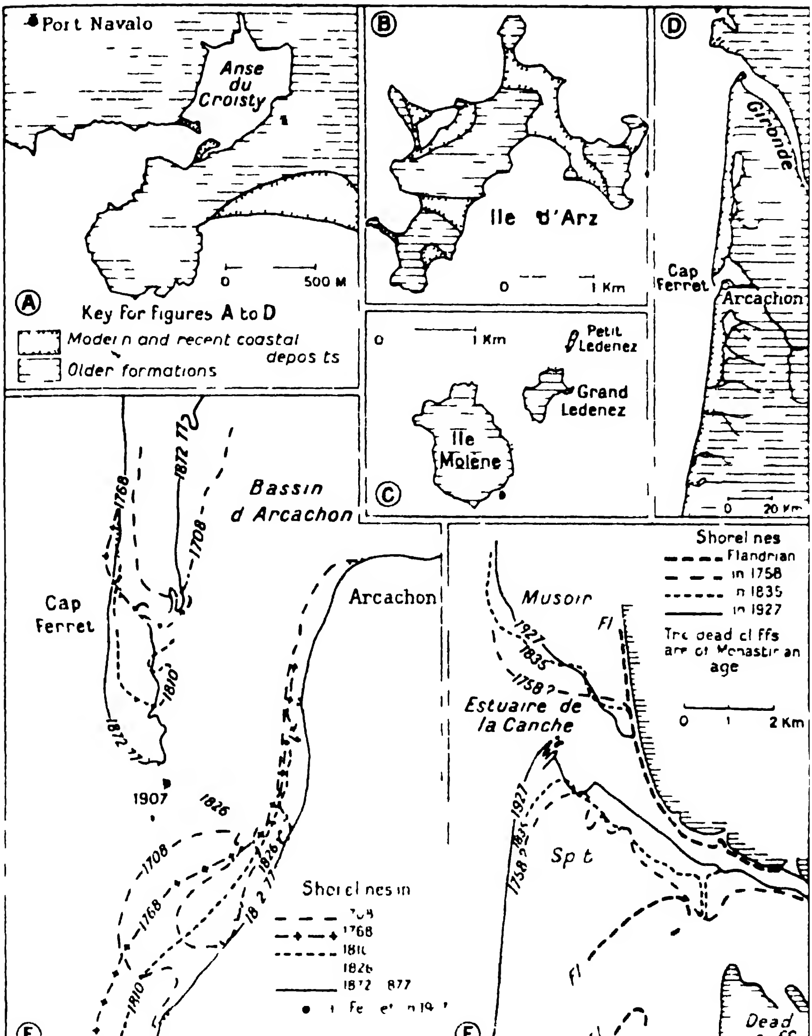


FIG. 6 TOMBOLOS, LAGOONS, AND SPITS

A Midbay bar in Croisty cove, near the entrance to the Gulf of Morbihan. B Ile d'Arz in the Gulf of Morbihan this is formed of islets connected by recent and Flandrian tombolos (after Guilcher, 1948). C. Ile Molene in western Brittany the tombolos connecting the islands are submerged at high tide (after Collin, 1936). D Connected lagoons on the Landes coast of Gascony E. and F Spit extension and erosion on the opposite shore (musoir) at Arcachon in the Landes and on the Canche estuary in Picardy (after Duffart, 1908 and Briquet, 1930) Note the temporary regression at Cap Ferret between 1826 and 1872–7.

Much more common are the spits of sand or shingle which extend across the mouth of a bay or an estuary from one side or the other. These are bay-mouth bars: the Germans call them *Nehrungen* and Briquet terms them *pouliers* in Picardy. The bay-mouth bar isolates a sheet of water which may finally become a lagoon; the connecting channel has a width depending upon the volume of water passing through it. Where the dominant swell is oblique to the coast and sets up a longshore drift, the bay-mouth bar grows in the direction of this drift (towards the north on the coast of Picardy), while the opposite bank, the *musoir* of Briquet, is worn away, so that the channel is constantly deflected in the direction of the drift and the mouth aligns itself parallel to the coast: for example, the Langue de Barbarie at the mouth of the Senegal river, which generally grows towards the south but is breached from time to time by heavy swells; the bay-mouth bars of Arcachon and the *Nehrungen* of Poland may be included in this class (Fig. 6 E and F).

In front of other coasts spits of sand originally unconnected with the shore are developed: they are called offshore bars (*crêtes d'avant-côte émergées*). Often they have a rock foundation which may be only a reef and not a true island, and from which two V-shaped points of sand or shingle develop, as at Mousterlin, south of Quimper (Fig. 7 A and B). But such a foundation does not always occur. The finest offshore bars are those which fringe the south-east coast of the United States and form Cape Hatteras. We may also quote those of Texas, and those of the south-east coast of Iceland described by Lewis. It is difficult to see how offshore bars without any foundations are formed, unless on originally submerged ridges. It is certain that at least those formed in front of deltas (p. 115) are built up in this way. The lagoons which they cut off tend to develop into marshes, which will be discussed later on.

Offshore bars are usually driven shorewards, but if they block important streams, one or more narrow channels must remain open. The channels through the bar may change, as at Mousterlin, where one of them has rapidly approached the point of the V since 1848 (Fig. 7 A and B), and at Katama Bay, New England (Johnson, *New England*, pp. 452–4).

Offshore bars can therefore be transformed into beaches blocking lagoons, but lagoons are not necessarily formed in this way; and, as Zenkovitch says, it is not true to assume that bars with lagoons behind them are characteristic of shorelines of emergence as they are just as

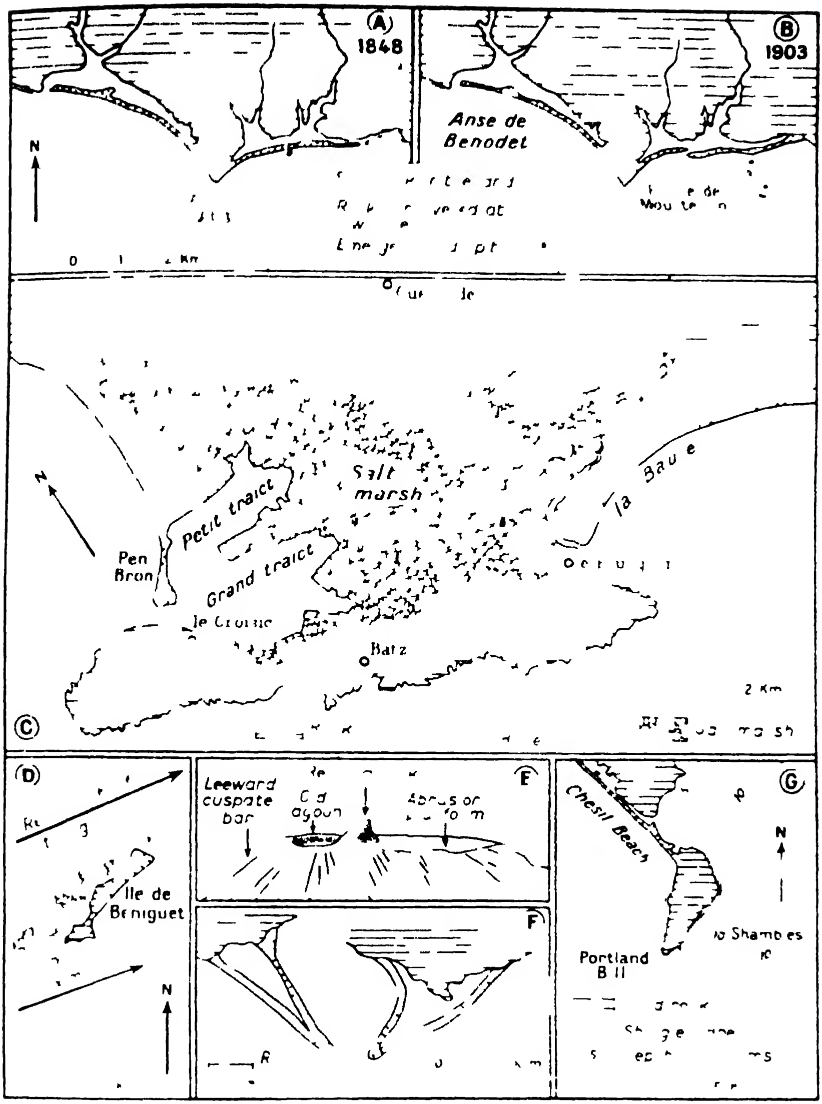


FIG 7 DIFFERENT TYPES OF SPITS

A and B V-shaped spit at Mousterlin, south-western Brittany, barring estuaries note the change in position of the eastern gap (after Guilcher, 1948) C Double tombolo at Guerande (La Baule and Pen Bron) D Leeward spit at Beniguet, Finistere (after Guilcher, 1950) E Cup Butte in the former Lake Bonneville, U S A, with cuspate bar enclosing a lagoon (after Gilbert) F Similar forms in Lake Bonneville (after Gilbert) G The tombolo of Chesil Beach and the banner bank of the Shambles in Dorset.

common or even more common on shorelines of submergence: the lagoon behind the bar may never have been a part of the sea.

The tombolo is another classic coastal accumulation form. This term, introduced into international geographical literature by Gulliver, was applied by him to a spit of sand or shingle which joins an island to the neighbouring coast. There are grounds for not changing the meaning given to the term by Gulliver and therefore not applying the term, as is sometimes done, to the whole complex of the spit and the old island. Simple tombolos are found, as at Quiberon (Fig. 28 B) and at Portland (Fig. 7 G); double tombolos as at Bourg-de-Batz (Loire Inférieure) (Fig. , C) and at Giens near Toulon. At Batz and Giens, the two tombolos are not equally developed; at Batz the western tombolo, Pen Bron, is not completely joined to the island. The triple tombolo at Orbitello in Italy is a special case. Certain tombolos are partly submerged at high water and dry at low water. In the Molène archipelago in Finistère, certain islets are connected twice a day to the larger islands and are called in Breton '*ledenez*' ('extension of the island') (Fig. 6 C). In places several islands are interconnected by tombolos; the celebrated example of this is Nantasket Beach in Massachusetts where the islands are drumlins some of which have been eroded away but have left traces in the outline of the composite tombolo.

Simple tombolos, or even double ones like that at Giens, may with few exceptions be explained either by the refraction of waves behind the island, if the submarine contours have an appropriate shape, or by their diffraction on each side of the island and the deposition of materials at the meeting point of the two diffracted and weakened sets of waves.

Triangular headlands of sand often spring from an island or the coast of the mainland. Depending upon their evolution, they may or may not dam up lagoons between their two flanks. These features may have been formed in several ways, other than by development from offshore bars.

(*a*) When they are connected to an island they resemble a comet's tail in outline and are often developed parallel to the direction of the dominant swell in the shelter of the island. This principle, formulated by Schou, is exemplified in the islands of Anholt in Denmark and Beniguet in Finistère (Schou, 1945 and Guilcher, 1950) (Fig. 7 D). Other examples are the headlands of sand sheltered from the westerly swell in the east of the island of Groix, and the pointe des Galets in Réunion sheltered from the south-east trades. A remarkable example was quoted by Gilbert from Cape Butte in the former Lake Bonneville in

the United States: a triangular headland surrounds a lagoon, while the direction of the dominant swell is betrayed by the development of an abrasion platform on the opposite side of the island (Fig. 7 E). This law applies to constructional forms having only one point of contact with the land. They are explained by the diffraction effects already discussed in connexion with tombolos.

Completely submerged banks may occur some distance offshore and were called banner banks by Vaughan Cornish. They occur behind headlands which weaken the swell and promote deposition. In the English Channel there are two remarkable examples: the Shambles Shoal east of Portland Bill (Fig. 7 G) and the Skerries Shoal to the east of Start Point.

(*b*) Certain tombolos of the comet's tail type, which may be called relict bars, originated like the preceding type in the shelter of an island; but the island has disappeared through erosion, and the spit alone survives resting on the abrasion platform of the island. This type, called flying bars by Gulliver, has also been studied by Johnson and by Nichols. Good examples of them are found in the harbour of Boston in the United States where the former islands were drumlins, which were easily worn away. With three dominant wind and wave directions flying bars with three tails may occur, e.g. in Boston harbour.

(*c*) Triangular headlands also result from two dominant swells more or less opposed to one another. Dungeness is an excellent example (Figs. 8 B and 27 B). At this spot there are two dominant swells, the main one from the south-west, the subsidiary one from the north-east. These two opposite swells explain the growth of the long shingle headland out into the Straits of Dover. Gulliver was certainly mistaken when he explained Dungeness by the meeting of tidal currents. The inequality of the forces acting upon the ness leads to erosion on the south face and a migration of the far point towards the east-north-east: the reasons for this are examined below.

(*d*) Triangular points of sand which occur in front of islands do not belong to type (*a*) and are incipient tombolos or, as Gulliver called them, uncompleted tombolos, e.g. Pen Bron (Fig. 7 C).

(*e*) Others have an uncertain origin: for example, several of those of Lake Bonneville (Fig. 7 F). It is possible that they belong to one of the preceding classes or they may be offshore bars, but for some reason or another their origin is uncertain.

Growth and development of spits: the fulcrum or dead point. The lengthening of spits or bars having free or unattached ends is generally

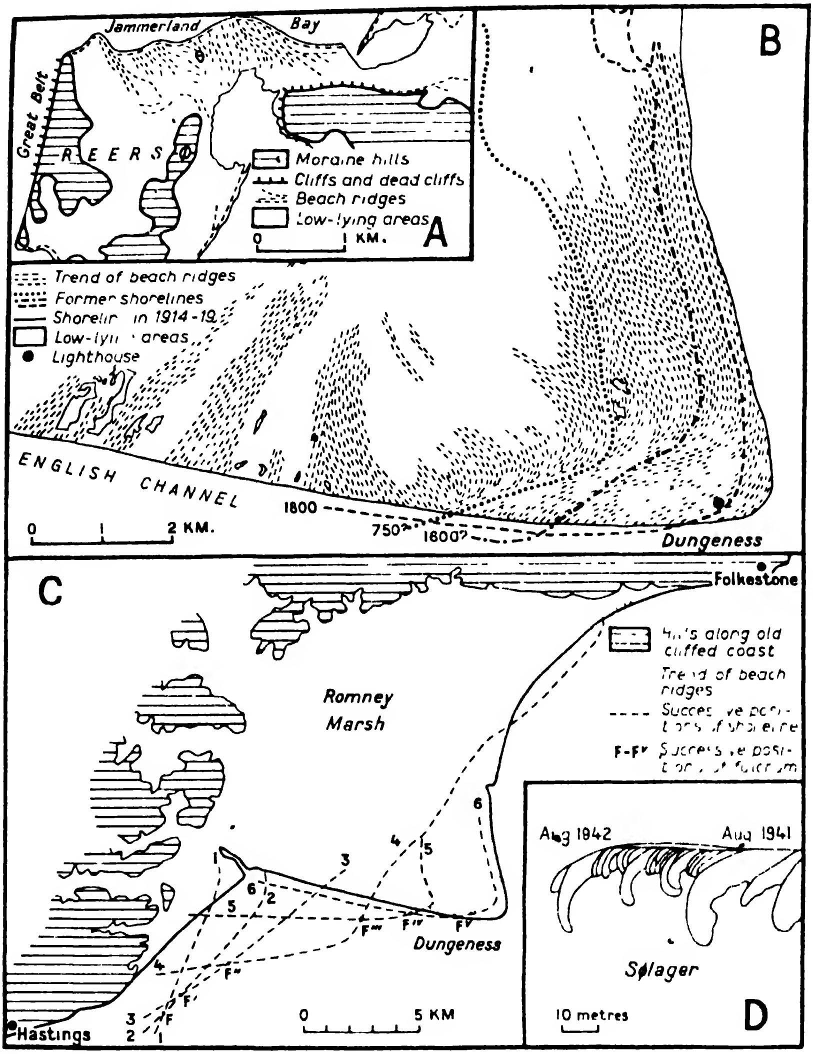


FIG. 8 COASTAL FEATURES FORMED BY SERIES OF BEACH RIDGES

A. General features of Reersjo, Zealand, Denmark (after Schou, 1945). B and C. The evolution of Dungeness (after Lewis and Balchin, 1932 and 1940). D. Spit with recurved laterals at Sjolager, Zealand, Denmark (after Schou, 1945).

brought about by the addition of hooks or lateral ridges. Such forms may be called hooked bars, or recurved spits. Examples are the modern and late Pleistocene spits of the Picardy coast (Briquet), Cape Cod (Davis), Blakeney Point and Scolt Head Island in Norfolk (Steers), and the numerous bars of the Danish and German Baltic coasts (Schou and others). The lengthening is brought about by longshore drift which is generally caused by the oblique incidence of the waves; recurved ridges may be due to the refraction of the waves at the distal end of the spit.

Widening of the spit is connected with its lengthening. During each period of strong winds and swell, provided that there is an abundant supply of material, a new storm beach may be built in front of the existing ones. This ends in a new recurve beyond the previous ones: there are, therefore, extensions of the spit simultaneously towards the open sea and also parallel to the shore. The result is the building up of a large number of parallel ridges. In certain cases studied by Schou, for example at Sjolager in Zealand (Fig. 8 D), more than twenty ridges and recurved laterals are built in a year; but the number of major ridges is very much smaller.

This building up, however, does not widen the spit throughout its length. The proximal end, or root, is often attacked and a general retreat of all the old part of the spit ensues. There is therefore a point, called the fulcrum by American authors, and the dead point by Briquet, which is displaced in the direction of movement of the spit and separates the part undergoing erosion from that which is being extended by the formation of new ridges. The result is that the inner and older ridges are cut by the new shoreline at an angle which is greatest for the oldest ridges and may even exceed 90 . This has happened, for ample, at Dungeness where the south coast is eroded and in consequence cuts across the old ridges: the obliquity is noticeable at the point and increases from east to west (Fig. 8 B and C). The migration of the fulcrum or dead point is illustrated at Dungeness, at Cape Cod, the Swinemünde spit, at Orfordness near Ipswich, at the Crumbles near Eastbourne, on the complex bay-mouth bars of the Picardy rivers, i.e. the Somme, Authie, Berck, and Canche, and in Denmark at Reersjo and Sevedjo (Fig. 8 A), and at the Pointe de la Coubre in front of the Gironne estuary. Easily eroded morainic coasts supply a wealth of material for marine construction and are most favourable for the multiplication of these delicate structures, a certain number of which, like Dungeness and the bay-mouth bar of the Somme, are true wonders of nature.

The reader will have noticed that nearly all beach formations have been attributed to waves or the longshore drift which results from them. Towards the end of the last century, numerous authors, such as Gulliver, believed in the great efficacy of tidal currents. Under the influence of Johnson, Lewis, and many others general opinion has now swung round and it is remarkable that modern ideas are often the same as those held by de la Beche in England and Élie de Beaumont in France in the first half of the nineteenth century.

COASTAL DUNES [1]

The frequency of dunes on the coasts is connected with the supply of sand from the beaches, with the absence of obstacles in the sea, thus allowing the wind to blow more strongly than elsewhere, and with the absence of vegetation in that part of the beach from which the sand is derived. However, a certain type of vegetation favours the formation of coastal dunes.

On sandy beaches of considerable width at low water, ephemeral dunes may be formed when the wind is strong, but these are destroyed by the rising tide. These are tongue-shaped dunes, or little barchans with the convexity facing upwind and the points tapering downwind. Examples are to be found at Blakeney in Norfolk. Somewhat more stable dunes of the same type occur on the upper parts of beaches which are only covered by the sea when there are storms or very high tides: they can be seen on the beach of Irhzer Amayane between Cape Rhir and Agadir in Morocco. But permanent dunes can exist only beyond the reach of the sea. They are in fact connected with the growth of land vegetation which fixes the sand brought from the beach by the wind. On the Atlantic and Mediterranean coasts of Europe two plants play an essential constructive part in this: *Agropyrum junceum* and *Psammia arenaria* (marram), the former being slightly tolerant of salt, but not the latter. Marram, which replaces *Agropyrum* under suitable conditions, is the more important of the two. It requires a continuous supply of fresh sand for its growth. In regions outside Europe, other plants may replace it. In this way small lenticular heaps of sand are formed rather like the *nebkas* of the Sahara. These coastal *nebkas* occasionally reach a height of several metres, for instance between Mogador and Cape Sim in Morocco. Usually they tend to unite into a long dune with an irregular crest bordering the shore. This happens on the majority of

[1] The general principles of aeolian accumulation are not studied here.

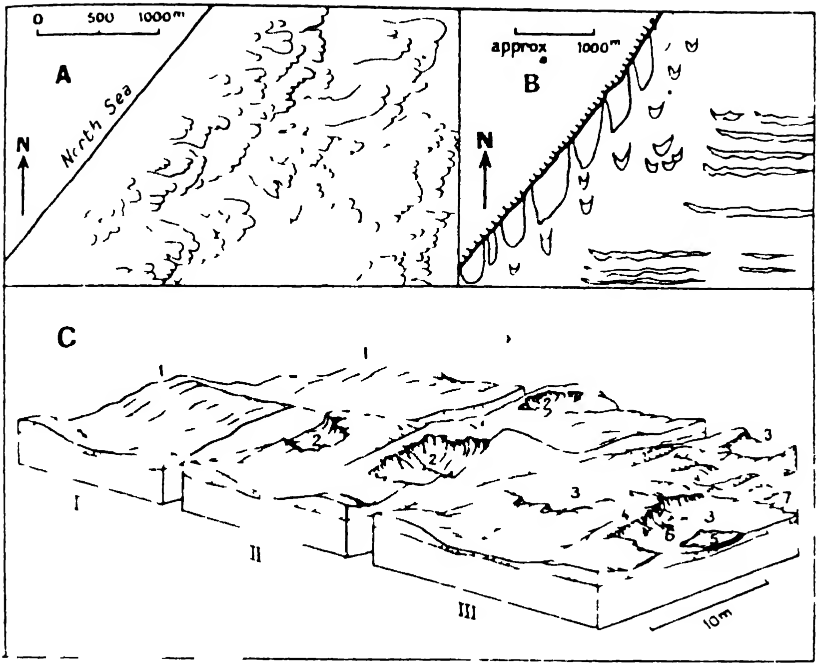


FIG 9 COASTAL DUNES

A Parabolic or 'rake' dunes formed by westerly winds north of The Hague, Holland (after Faber and J Hol)

B. Dunes south of Agadir, Morocco, formed by the trade winds deflected by the Atlas Mountains. Deflation occurs on the crest of the dune-sandstone cliffs, and the sand reforms as parabolic dunes, then as wave-like dunes (after Guilcher and Joly based on oblique aerial photographs).

C. Evolution of transverse dunes (after Schou) 1 Wave-like transverse dunes 2 Blow-outs 3 Deflation areas 4 Residual dunes 5 Pool. 6 Talus. 7 New, reconstituted, dunes

European coasts notably in Gascony; but it is not universal, and in Morocco, for example, the coastal *nebkas* appear quite stable and occur on the seaward side of extensive areas of barchans bare of vegetation (pl. IV). But barchans are not specifically coastal dunes. According to Aufrère, the barchan characterizes regions with little or no plant cover: it is found therefore in coastal areas only in arid or semi-arid districts or high up on unstable beaches. The horned dune of humid regions may in fact be of the parabolic type (see below).

In temperate regions when the stage of the frontal dune is reached, the plants instrumental in the initiation of dunes tend to give way to plants which fix them: *Carex arenaria*, *Crucianella maritima*, lichens, etc. Unlike marram these plants do not require supplies of fresh sand to keep them alive. But the frontal dune, although established, remains weak. Winds can breach it: the *caoudeyres* of the Gascony dunes are formed in this way. Such breaches are known under different names on the coasts of the North Sea and Picardy (Fig. 9 C) and as blow-outs in the British Isles. Caoudeyre is a dialect word meaning cauldron. They are semi-circular hollows. The roots of the plants are exposed but hang from the edges of the hollows and prevent the sand from collapsing. The blow-out is enlarged by wind action. It cannot be cut down below the water-table in the dune; if it reaches it, a pool or marsh appears. On the downwind side, the sand forms a mound of confused shape (Fig. 9 C).

The frontal dune may thus almost completely or entirely disappear, the eroded remains being known as *crocs* in Picardy. It is then replaced either by parabolic dunes or by lines of dunes parallel to the wind. Parabolic dunes, of which there are good examples at La Baule (Loire-Inférieure), between the Somme and the Authie, in Holland north of The Hague, and in Denmark, are crescentic in form like barchans but their convexity is downwind and the wings point upwind, perhaps because the sand is not so mobile in humid as in arid countries. The convex main part of these dunes is possibly derived from the shapeless mass of sand which is formed at the end of the blow-out; the wings are accumulations of sand moving along the sides of the blow-out. Series of parabolic dunes produce rake-like forms, for example at Cazaux, at Cape Breton (both in Gascony), and in Holland (Fig. 9 A). Dunes parallel to the wind are parabolic dunes with no central mass but only wings.

Different dunes in the same complex do not always travel at the same speed. Dunes may, therefore, overtake each other and unite. Examples of these are found on the south-east coasts of the Baltic (Hartnack).

IV Coastal barchans on top of fixed dunes orientated in the same direction, Mogador, Morocco. (*Photo. I.G.N.*)

Frontal dunes may re-form upwind of parabolic or 'rake' dunes and continue to feed them from blow-outs. Parabolic dunes or dunes parallel to the wind may form without frontal dunes, as in Morocco near Tiznit (Fig. 9 B). In this case it is possible that the supply of sand may have been guided by the underlying relief, at least near the sea.

Finally, the sand removed from the frontal dune may collect again behind in a dune perpendicular to the wind. Wave-like dunes are thus formed parallel to each other and separated by flat-floored troughs called *lettes* in Gascony, where such wave-like dunes are particularly well developed (Fig. 9 C). The largest dunes outside deserts, apart from those resting on rock foundations, are of this type. For example, the dune of the Pyla at Arcachon is about a hundred metres high but its exact height changes constantly. These dunes are the most suitable for transforming estuaries into lagoons as in Gascony.

All moving dunes have asymmetrical outlines as do current ripple marks: the windward slope is gentle, and the leeward slope steep corresponding to the angle of rest of the sand (about 32°). All moving dunes may be subject to blow-outs.

The orientation of coastal dunes depends upon the direction of the most effective wind. The latter is often an onshore wind perpendicular to the coast, as such a wind is the least likely to be retarded but this is by no means an absolute rule, as an oblique wind may be sufficiently regular and strong to prevail over the perpendicular wind. Schou mentions blow-outs at Skagen in Denmark, the axes of which make an angle of 40° with the shore, while the axes of the parabolic dunes of La Baule (Briquet) are almost parallel to the beach. On the coasts of East Pomerania and of the Kurische Nehrung, the prevailing wind is slightly north of west. In both areas there are dunes perpendicular to the wind, but their relation to the shoreline is very different since the coast is east-north-east–west-south-west in Pomerania and north-north-east–south-south-west on the Kurische Nehrung (Hartnack). According to Kuhnholtz-Lordat, most of the blow-outs on the coast of the Gulf of Lions are hollowed out by the offshore mistral. In Morocco the orientation of the dunes is generally governed by the trade winds: the result is that at Cape Sim where the coast changes from north-east–south-west to west–east, dunes almost parallel to the trade winds and the coast are formed on the north-east–south-west shore, and are then brought to an end on the west–east shore: in this case it is the dunes which feed the beach and not the other way round. Similarly the port of Arcila in the northern part of Morocco has been recently silted up by the action of

the offshore north-east winds which have built a veritable 'aeolian tombolo' in the harbour after the unfortunate construction of a breakwater preventing the swell from clearing away these deposits.

Waves frequently erode coastal dunes. They cut a cliff in them either temporarily in a storm or permanently. When the sand is wet, the cliff is vertical. If the sea does not reach it for some days and if there is no rain the cliff is covered with dry sand falling from above and its surface assumes the angle of rest of the material (32°). The recession of the shore as a whole and the progression of dunes inland are two independent phenomena. They occur together on the coasts of Gascony, but this is doubtless a coincidence.

Dunes often bury pre-existing relief, and give the impression of a great thickness of sand although the latter may form a covering of only a few metres. Most of the dunes on the coast of Brittany which exceed 15 m. in height are of this type, as well as those reaching 62 m. at Carteret and 80 m. near the Anse de Vauville in Normandy. The dune-forms then reflect the underlying relief.

A great number of the coastal dunes of humid regions are fossil dunes. In certain cases this is clear from the fact that they are perched on cliffs and completely deprived of sand supply: for example those on the coast west of Quiberon, where they have been abandoned during the movement inland of the dune complex associated with the tombolo of Penthièvre (Fig. 28 B). More often the decay of the dune results from a change of climate which has caused its complete fixation.

In dunes which have reached a certain degree of stability, a bed of sandstone (alios) is common. This layer, which in certain cases is the result of podsolization, has the effect of preventing the dune as a whole from resuming its movement. But in humid temperate regions, and when the sand is non-calcareous, this process scarcely ever involves the whole dune. The majority of the calcareous sand dunes of the Mediterranean and arid regions show different features. In the islands of Hyères, in the Lebanon, Morocco, and in west and south Australia there are a number of examples of large dunes fixed by calcareous sandstones of Pliocene and Pleistocene age. Moreover, in certain of these countries the surface of the fixed dune is covered with a very hard crust. As several generations of dunes have been successively consolidated in this way in the course of a progressive emergence, a piling up of dunes to a considerable thickness has occurred in places, each dune overlapping the one previously affected (pl. IV). Dunes converted into sandstone may also form parallel lines as in the Casablanca–Rabat

region and in south-east Australia. The lowest and most recent consolidated dune is today attacked by the sea; when only the outside slope of a consolidated dune parallel to the coast is eroded, a hog's back is formed. A more advanced state leads to the development of a form simulating the barrier reef of the coral regions. In other places, they are at present being covered by sand. These different aspects are well represented on the Moroccan coast. Coastal dunes may, in this way, preserve for a very long time the pattern the wind has given them, and so form the resistant skeleton of the shore.

C. ESTUARIES, MARSHES, AND DELTAS

Estuaries and Marshes. The word estuary comes from the Latin *aestus*, meaning tide. In France it has a popular synonym, *étier*, which is widely used in the region of the lower Loire and the east of Morbihan. The two words denote that part of a river system (pl. V) where the tide, either the salt tide or the dynamic tide, and its currents make themselves felt.

Tidal marshes (Fr. *marais maritimes*) are low tidal zones approximately at sea-level. They are not necessarily found at the mouths of rivers. Where the river ends in coastal flats, marshes and estuaries are found together, as at the mouths of the Loire, Gironde, Dollart, Weser, and Elbe. Nearly everywhere, moreover, small streams flow into the marshes.

Marshes are widespread on all the low coasts of the world: for example in France, there are the marshes of Carentan, Croisic, and Batz, and the Breton, Poitevin, and Saintonge marshes; on the North Sea there are the Flemish, Dutch, and Frisian marshes and in England the fens and marshes of Norfolk and Suffolk; in North America they occur in Carolina and Georgia; in tropical countries, marshes are found on the Gulf of Guinea, notably in the Cameroons, Gabon, Portuguese Guinea, and on the river Casamance. Marshes occur in front of certain regions of marked relief, for example Morfa Dyffryn and Morfa Harlech in Cardigan Bay, *morfa* meaning marsh in Welsh. The marshes have often been reclaimed as in the Low Countries, Flanders, the Fens, and East Friesland. But in nearly all marshes unenclosed parts remain, which may be either near the sea like the Wadden Sea in Friesland and the bay of Veys in Normandy, or towards the interior behind the well-drained polders, where the level is a little lower, for example, the damp marshes of Poitou, Dol, and the bay of Fundy.

From the point of view of origin, there are at least three kinds of marsh; those which are in the sheltered part of an estuary and are formed by silting at the sides of the estuary, as in the Gironde; those which spring up behind a sand or shingle spit, e.g. Romney Marsh behind Dungeness and the marshes behind Blakeney and Scolt Head Island in Norfolk; those which are formed at the head of a bay into which no large river flows and which is unprotected by a sand spit, e.g. Carentan, the Poitevin Marsh, and the Fens.[1]

Marshes and estuaries are drowned land areas which were later silted up, the flooding being due to the Flandrian transgression. But occasionally isostatic movements have intervened notably in glaciated regions like east Scotland. The detailed history of marshes is always extremely complicated, but its interpretation is facilitated by the existence of datable deposits. Deposition gives estuaries and marshes their present form. Usually it begins at a depth of 30 m. at the mouth of the estuary or offshore; sometimes it is less, rarely more, e.g. 43 m. at the mouth of the Gironde, 117 m. in the Mahajamba estuary, and — 67 m. in that of the Majunga in Madagascar. The longitudinal profile of all estuaries is very irregular: in narrows, tidal currents hinder sedimentation and form scour holes; in front of the mouth of nearly all estuaries the depth is small and hinders navigation. The scour holes may be due, not only to an absence of sedimentation, but even to a real deepening in the solid rock when the latter is not very resistant. Thus the holes, 55 and 59 m. deep on the Scheldt in front of the Zieriksee lighthouse (Fig. 11 B) and Terneuzen, behind depths of less than 10 m. at the entrance, are certainly due to erosion in the Tertiary beds, the remains of which are thrown up on the shores of Walcheren. In holes in hard rock like those of the Goulet of the Rade de Brest (50 m.) and Morbihan (31 m.) erosion of this sort, on the contrary, is most unlikely.

Sedimentation in marshes and estuaries differs in certain respects. It is more uniform in estuaries, though not always entirely so. Sometimes there are coarser beds especially towards the base: for example in the Gironde, the Charente, and the Vilaine. The upper part of the marsh is nearly always muddy, at any rate away from the beds of tidal channels which are usually sandy, gravelly, or stony. The only exceptions are little estuaries near large sandy beaches or dunes, e.g. Laïta in Morbihan; Le Conquet and Quillimadec in Finistère. But even in

[1] Marshes may also develop on open, shallow-water coasts, such as that of north Norfolk (Translator).

V Estuary of the Wadi Dra, Morocco
Photo Aéronautique Navale

these the proportion of sand becomes less towards the head of the estuary.

These muds, studied by Glangeaud, Bourcart, Francis-Boeuf, Rivière, and Berthois, are found not only in the estuaries of western Europe and Morocco but throughout the world e g in parts of the African and north Pacific coasts

The colour of the mud varies It may be red as in the Bay of Fundy, but is generally black or dark grey Breton mud is blacker than the mud of the Charente and Guinea Mud contains a variable percentage of sand but consists mainly of precolloids and colloids, especially clay minerals, possibly kaolinite, illite, and montmorillonite It also contains iron, iron oxide is the cause of the redness of the mud of Fundy Iron forms the cementing material, together with organic matter The latter, forming on average 8–10 per cent of the total, appears to come from plants and animals, some of which live in the mud some is derived from plankton brought from elsewhere Mud is soft in the sense that one sinks in deeply when walking There is even a risk of sinking completely, but the depth to which one may sink varies with the estuaries and is very small when there is a large percentage of sand Mud is plastic and rigid, and keeps the shape impressed upon it, footmarks remain quite clear in it It has the property of liquefying when patted

It is easy to assume that mud is formed by the finest sediment brought down by the river and deposited at the top of the marsh, the sediment is fine because the river downstream cannot transport coarser material But scientists have not come to any agreement on this point. One fact is certain, mud is deposited in places where tidal currents are strong Its origin is controversial, and perhaps varies from estuary to estuary According to the sedimentologists of the Low Countries, the deposits of the Dutch estuaries are derived from the sea, and not from the rivers this is explicable by the fact that on the floor of the estuary the current is upstream (p 22) For the Loire, however, Berthois comes to the conclusion that mud does not come from the sea because the amount of material in suspension is very small near the sea

The same author has clearly established that, in the Rance estuary in north-east Brittany, mud has been supplied by the fine periglacial deposits washed at high tide from the banks of the estuary itself, and the same thing also holds good for other small estuaries The crucial fact is that, in all estuaries, the amount of sediment in suspension is far greater than both that in the river upstream and that in the sea It is this which makes the problem of origin so difficult For example, above

the estuary the Garonne carries, according to Glangeaud, 80,000 tons of sediment in suspension during the period of one tide; this figure increases to 4,500,000 tons in the Gironde, between Saint-Christoly and Blaye, and falls to 70,000 at the mouth. This mass of mud in suspension oscillates up- and downstream with the tide. According to Berthois, in the Loire this to-and-fro movement only occurs in ordinary and neap tides in summer; but, during the floods of the river in winter, a great quantity of sand coming from upstream is deposited in the lower part of the estuary; on the other hand, during spring tides, a part of the mud in suspension is deposited far upstream in Nantes harbour at high tide, and another part is carried outside into the sea at low tide, where it is distributed over the Breton marsh.

The nature of the deposition itself is also complex and controversial. According to Francis-Boeuf the mud in suspension is not deposited by flocculation brought about by the salts present in sea-water: this would explain the agglomeration of the particles but not their deposition. However, Berthois, Chatelin, and Marcou have shown that deposition in the Loire is accelerated by salinity and also by heat. According to Francis-Boeuf, mud is deposited on pre-existing mud-banks at the side of the channel during the ebb-tide, in the same way that scum sticks to the side of a bath when it is emptied, firmly enough not to float off when the bath is refilled. But Berthois thinks, on the contrary, that deposition principally occurs during slack water. Glangeaud adopts the same explanation as Francis-Boeuf for lateral banks, which Francis-Boeuf thinks are the only ones: the channels are fixed and free of mud. In the Charente channel at Taillebourg even limestone pebbles are found. In Zealand the beds of the estuaries are sandy, and the mud, according to marine charts, is deposited at their edges. The same is true of the Breton estuaries, but only in a general way: in some sections with a gentle slope, mud is found in the middle, notably in the Aber Wrac'h. Glangeaud and Rivière stress the importance of mud-banks in the middle of the Gironde and in the Lay (Vendée). According to Glangeaud such sedimentation cannot be produced in the same way as that on the lateral banks, since the channel never becomes dry: it is caused by eddies with a horizontal axis. This author explains their formation as follows (Fig. 12 E): in a straight channel the maximum velocity is at the centre near the surface, and two maxima of turbulence occur one-third and two-thirds of the distance across the stream half-way between the surface and the bottom. A descending current forms under the line of maximum velocity between the two maxima of turbulence,

and supplies the median bank. Once this latter is formed, it divides the channel into two branches, in which the same phenomena are reproduced so that there will be a tendency towards the formation of new median banks in each of the channels. In other words, the parallel median banks will multiply until an important change of outline is produced.

But is mud still being deposited in estuaries? Measurements made up till now have chiefly concerned marshes proper. In estuaries certain mud-banks seem to have reached a state of equilibrium. The estuary of the Charente, for example, is stable in spite of the great turbidity of its waters.

Yet, if one makes a hole in a mud-bank, this hole fills up very quickly. Dock basins which are excavated in the banks of an estuary must be closed by a lock or they will fill up with mud. In 1944 the Germans blew up the locks of the Rochefort basins which are connected to the Charente. The result was the deposition of 7 m. of mud in these basins between 1944 and 1947. The deepening of channels in estuaries below their natural equilibrium to allow the passage of large ships is indeed a Herculean task.

Soft mud areas are called *slikke* by the Flemish and Dutch, and mud-flats by the English. They are not completely flat in spite of the English expression: they may be examined in both estuaries and marshes, but we must first present an outline of sedimentation in the latter.

Sedimentation in marshes is variable. Apart from mud like that of estuaries, sand and both salt-water and freshwater peats are found.

The deposition of mud is the first stage in marshes enclosed by a spit, as, before the spit formed, these areas were ordinary beaches open to the sea. It may go on a very long time in the mud-parts of the marshes, as for example at Blakeney in Norfolk (Fig. 10) or in the Wadden Sea (Arenicola sands) (Fig. 12 A and B). It may even continue to cover the whole of the marsh, as in those of Cardigan Bay. But usually, in spite of transitory and local recurrences, it gives place to other types of sedimentation when the marsh is mature. At that stage finer material is deposited usually at the edges of the marsh. *Tangue* is deposited on the marshes of north-east Brittany: it is a predominantly sandy sediment, with some silt, an almost complete lack of clay, poor in organic matter, and remarkable for its high content of lime (from 25 to 61 per cent.), its silvery-grey colour, and its permeability. *Terre de bri*, deposited on the marshes of west central France, is bluish, and contains on the average, in the Poitevin marsh, 75 per cent. sand and fine sand, 15 per cent.

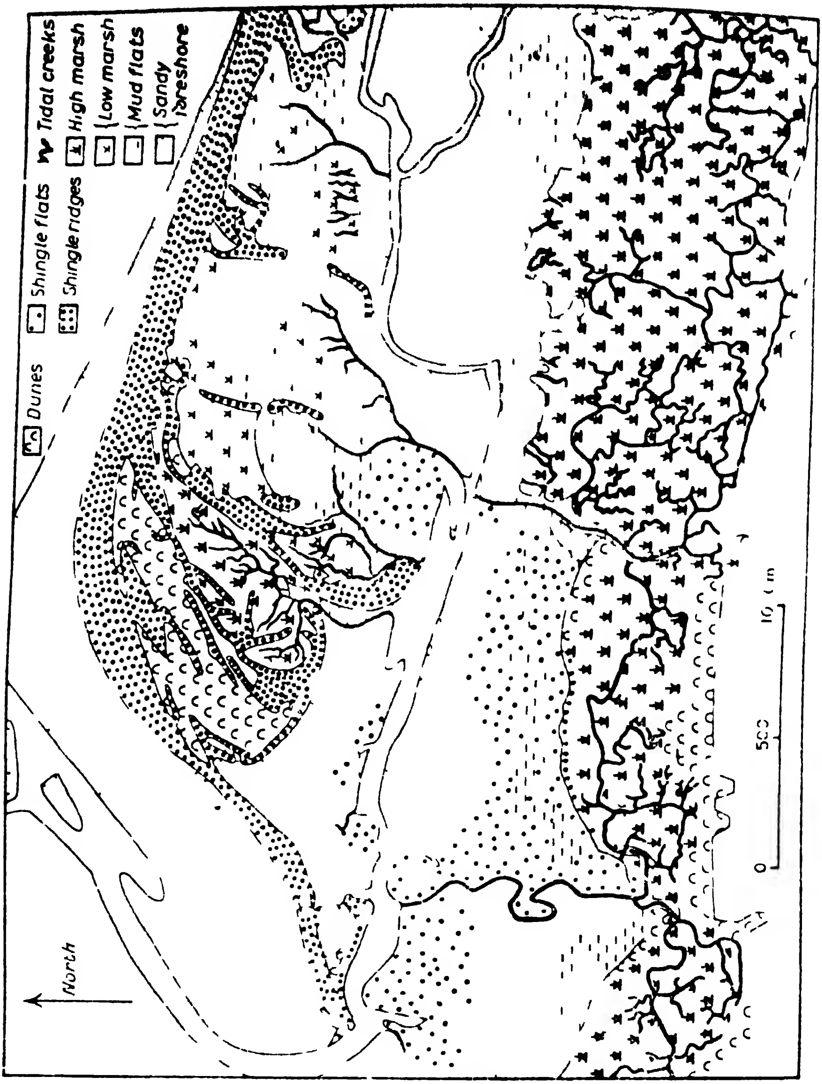


FIG 10 BLAKENEY POINT, NORFOLK (AFTER STEERS)

Recurved laterals, dunes and marsh sand in the middle and mud at the edges

still finer material, and 8·5 per cent. of lime. Pure mud, of which pre-colloids and colloids form 50–90 per cent. of the total, may also be deposited. These deposits may constitute the whole of the deposits, with no pure sand underneath. A film of mud may form on the sand, and may be eroded or thickened. But this mud acquires great stability once it is colonized by marsh vegetation.

Vegetation in estuaries, as in marshes, has a considerable morphological importance: certain species, such as *Spartina townsendii*, spread with amazing speed in the marshes of western Europe. They establish themselves first as tufts on the soft mud. Elsewhere *Salicornia* colonizes the mud-flats and causes its level to rise. These plants are followed by others such as the sea-aster, then by a grass, *Puccinellia maritima*. In tropical marshes the mangrove takes the place of grass. On the African coast mangroves begin to grow along the estuary of the Senegal in sheltered places and reach about 2·50 m. in height; further to the south they grow much higher. The roots of the mangrove are alternately covered and uncovered by the tide, and form an inextricable network. Mangroves do not necessarily require salt nor even brackish water to live; but a certain tidal range is necessary for the reproduction of the trees, which cannot occur in swamps which are always lying under water. Precise measurements would be necessary to discover whether sedimentation there is more or less rapid than on the mud-banks where no mangroves grow. Whether they are bushy or grassy, patches of vegetation gradually canalize the ebb and flow of the tide. As the vegetation cover becomes more complete, and as the mud dries, becomes less salt and more granular, the character of the marsh changes and only the network of tidal creeks remains clear of vegetation. These creeks are mainly produced by the upward growth of the marsh, and maintained by the constant action of the tidal waters in narrow channels, the banks of which may be some decimetres high in the lower part of a marsh and more than 3·50 m. in the upper parts (Fig. 10 and pl. VI A). Between these channels, the sides of which remain without plants, the vegetation thickens and develops as the marsh increases in height. The process is due partly to the deposition of mud at high water and partly, in some places, to the formation of peat derived from the vegetation of the marsh. The banks of the tidal creeks, often covered with *Obione portulacoides*, are steep or overhanging. Sections of the bank may slip down as the result of undercutting because the creeks develop beautiful meanders, both in tropical mangrove swamps and temperate marshes. At the sides of estuaries conditions are exactly the same.

As the marsh evolves, it develops into a salt marsh (Steers) where flat surfaces dominate and the system of channels disintegrates. The old network of creeks is broken into narrow discontinuous sections which vegetation tends to fill by growing down from the sides. These are not very deep (20–30 cm.), but may retain water at low tide, although they no longer form an organized drainage system. Salt pans are about the same depth and round or oval in shape. Vegetation cannot establish itself here because of great variations of salinity,[1] due to the evaporation of stagnant water (pl. VI C). The upper part of a salt marsh is only covered at very high tide: it is, however, intersected by occasional creeks, connected with the general creek system, the depth of which varies with the height of the marsh; these are filled by the sea even at neap tides.

Seaward of the vegetated marsh the bare sand or mud areas are also intersected by tidal channels, the banks of which are not so steep as those of the marshes. They are subject to much more rapid modification, such as shifting and the cutting off of meanders. The channels are often of a special character, some being followed by the flood current, others by the ebb: the system of flood channels splits up and dies out on mud-flats or sand-flats; the flood passes over it as a sheet, again canalizes itself temporarily and so on. Exactly the same happens at the ebb. Examples are found in the arms of the Scheldt and the bay of Mont-Saint-Michel (Fig. 11 B).

Vegetation is very sensitive to variations of environment. In the inner parts of the marsh or estuary reached only by the dynamic tide, there is usually a growth of freshwater peat, produced chiefly by *Phragmites* and the water Iris. The building up of the marsh displaces the salt-loving vegetation, and hence the peat zone, seawards. But if a storm breaches the beach in front of the marsh there may be a regression with the formation of marine peat above the freshwater peat; such an arrangement may also result from a marine transgression, but a great thickness of marine peat resting on freshwater peat must be present for such a conclusion to be reached. Cases of this kind have been discussed by Johnson in New England (*New England*, pp. 540–5).

The evolution of marshes is further complicated by two phenomena: the compressibility of the peat and the migration of the enclosing coastal forms inland.

Peat shrinks very quickly to about three-quarters of its thickness, or even less, when it is either compressed or dried. Compression is produced

[1] Lack of drainage is probably a more important factor (Translator).

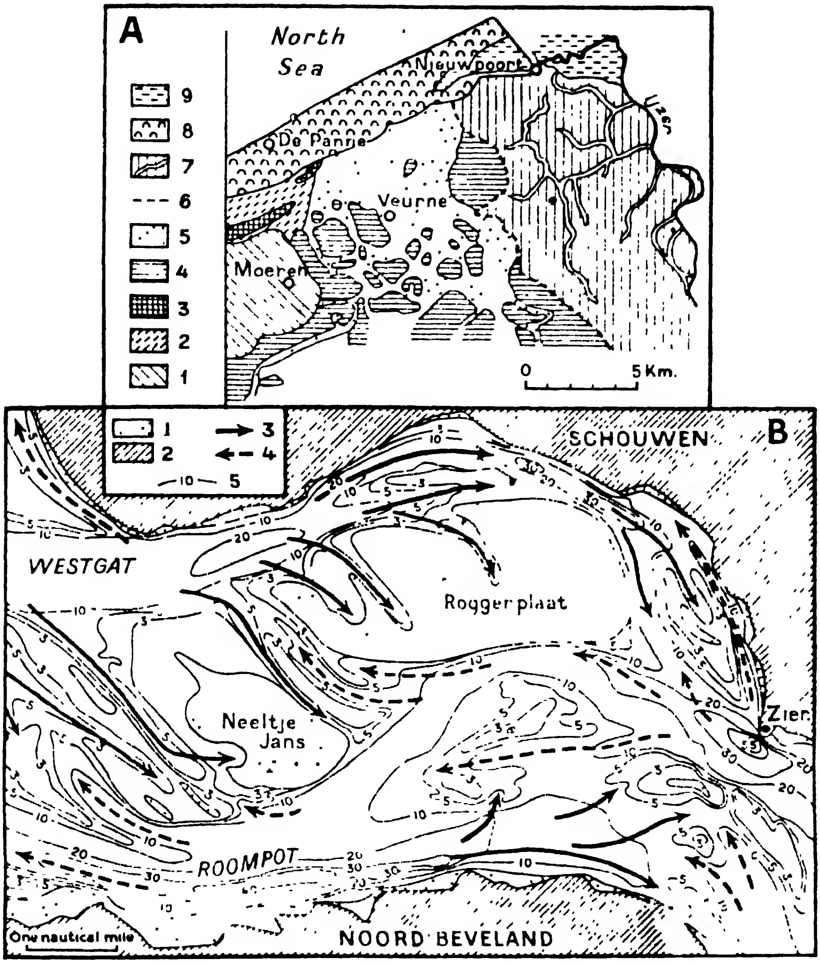


FIG. 11 MARSHES AND ESTUARIES OF FLANDERS AND ZEALAND (THE NETHERLANDS)

A. Furnes (Veurne) and Nieuport (Nieuwpoort) region (after Moormann). 1. Calais bed outcropping in the Dunkirk marsh after removal of peat. 2. Thin clay or sand cover on Calais bed. 3. Old dunes. 4. Isolated peat areas between old sandy channels, forming depressions due to compaction and inversion of relief. 5. Sandy channels or sand-flats. 6. Old sea wall. 7. Polders of the Yser estuary. 8. Recent dunes. 9. Modern polders in the Yser estuary.

B. Flood and ebb channels in the eastern part of the Scheldt estuary (after van Veen). Note the scour holes. 1. Tidal marshes and mud-flats. 2. Land. 3. Flood channels. 4. Ebb channels. 5. Contours of depths below average low spring tide level. Zier. Zieriksee light.

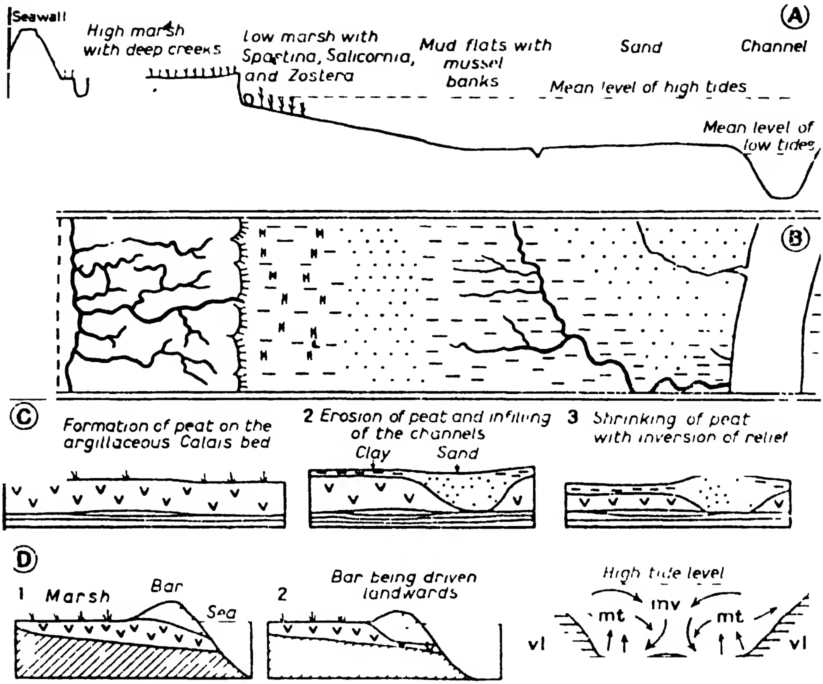


FIG. 12 SEDIMENTATION AND EVOLUTION OF MARSHES AND ESTUARIES

A and B. Diagrammatic section and plan of Wadden Sea area (after Van Straaten).

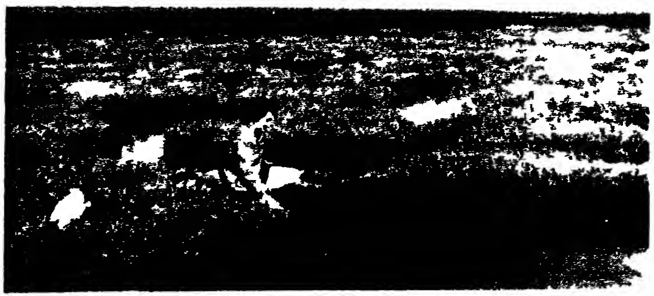
C. Inversion of relief caused by shrinking of peat in the plain of Flanders (after Tavernier).

D. Compression of peat by a bar being driven inland over it (after Johnson).

E. Formation of a median mud-bank in a straight channel (after L. Glangeaud): *mv* - maximum velocity; *mt* maximum turbulence; *vl* = lateral mud-banks; *s* sand; *vm* median mud-bank.

VIA Blakeney marsh Norfolk Tidal creek in high marsh with *Aster* and *Salicornia* *Obione portulacoides* at edge of creek

VIB Eroded high marsh Conquet estuary Brittany



VIC Blakeney marsh Salt pans

VID Pebbles from a depth of 200–300 m in the Saint-Tropez canyon Provence (*Photos* A *Guilcher*)

by new layers of peat, by deposits of mud driven on to the peat if the enclosing beach is broken, or by draining in the construction of polders. As a result, freshwater peat may settle well below the level of the sea without the sea-level itself altering. On the other hand, as peat tends to form in the inner parts of the marshes and only occupies the areas between the channels, even a relative drying up of the marsh may lead to an inversion of relief (Figs. 11 A and 12 C): the inner parts sink below the sand- and mud-flats. Channels above the level of the peat are very characteristic in many marshes, e.g. in Flanders, Holland, north-west Germany, the Fens of England, and the bay of Fundy. Thus marshes which show an increase in height seawards are produced. However, this classic explanation of the raised position of the channels has been contested in the case of the Fens and German marshes by Godwin, who produces arguments in favour of the idea that these channels have always been raised above the peat, and that this is due to the progressive building up of the banks by the tidal current circulating in the creeks. He recognizes, however, that the shrinking of the peat may noticeably accentuate the difference of level. The outcrops of old channels, known as roddons in the Fens, are often used for roads and houses.

The movement inland of the enclosing beach spit reduces the width of the marsh, and may allow the sea to penetrate farther. If the enclosing ridge happens to rest on peat it compresses it, and depresses it below its former level, so that after the ridge has moved further inland, the freshwater peat may lie below the level of the sea on the outer part of the beach. Again, such a state of affairs does not necessarily indicate a marine transgression (Fig. 12 D).

In many marshes on the Atlantic and Channel coasts of France, it is at present noticeable that the banks of the creeks not only collapse in the normal manner, but that the marsh with its cover of vegetation is being destroyed on its seaward side. It is reduced to a series of isolated hummocks actively attacked on all sides and known as *talards* in the St Malo district. The marsh of Conquet in Finistère is typical (pl. VI B). Is this phenomenon connected with a recent slight change of sea-level?

The presence of raised sections in marshes, never or hardly ever covered by the tide, might be construed in the same way. May they not have been built at a time of a slightly higher sea-level, and may the subdivision into hillocks not be the result of a slight regression? On the other hand, it can be argued from the fact that the hillocks are still covered at very high water that they are growing vertically as a result

of continual silting on their upper surface at the same time as they are being eroded at their edges. It may be that there is a cycle under present conditions; the mud of the eroded hillocks, apart from building up the surface of the hillocks, is deposited on the mud-flats which gradually evolve into marsh, which finally grows too high to be covered by any except very high storm tides which happen only once or twice a year.

More measurements, more levelling, more photographs from fixed points and more use of layers of sand, as employed by Niels Nielsen and Steers to measure the rate of sedimentation, together with systematic borings in marshes and estuaries, will doubtless solve some of the problems.

Deltas

A delta is a river mouth where alluvium accumulates instead of being redistributed by the sea. The delta, therefore, forms an extension of the land, sometimes of considerable size: 30 km. in the case of the Danube and 140 km. in the case of the Mississippi, which has the largest delta in the world. A delta need not have many branches. Some deltas, such as that of the Tiber, have only one branch, while the mouth of the Amazon, which is divided into branches, is not a real delta for the islands in it are not entirely formed of recent alluvia. The seaward projection is the main characteristic of deltas. However, most deltas have several branches.

Many deltas are to be found in seas without appreciable tides or in lakes, e.g. Po, Nile, Volga, Amu-Daria, and those of the Mississippi. But the great deltas of monsoon Asia are formed in seas where the tides are far from being weak. For instance, the tidal range is 4 m. at the delta of the Song-Koi. The absence of tides is certainly a favourable condition; but really large rivers overcome the unfavourable effects caused by the tides. It is not true that the greater the discharge of a river, the greater the chance of it building a delta in tidal seas. The Congo has no delta, probably because it ends in a deep submarine canyon difficult to fill, and possibly also because its load is not great at its mouth.

Even at the mouth of rivers without a delta, there is often a tendency for alluvium to be deposited. These embryo deltas are of two types: the submarine delta and the bar at the mouth of the estuary.

The Loire (Fig. 13 D) and the Vilaine afford examples of submarine deltas: the submarine contours indicate flattened submerged bulges with channels on either side. Ships have therefore to use these channels.

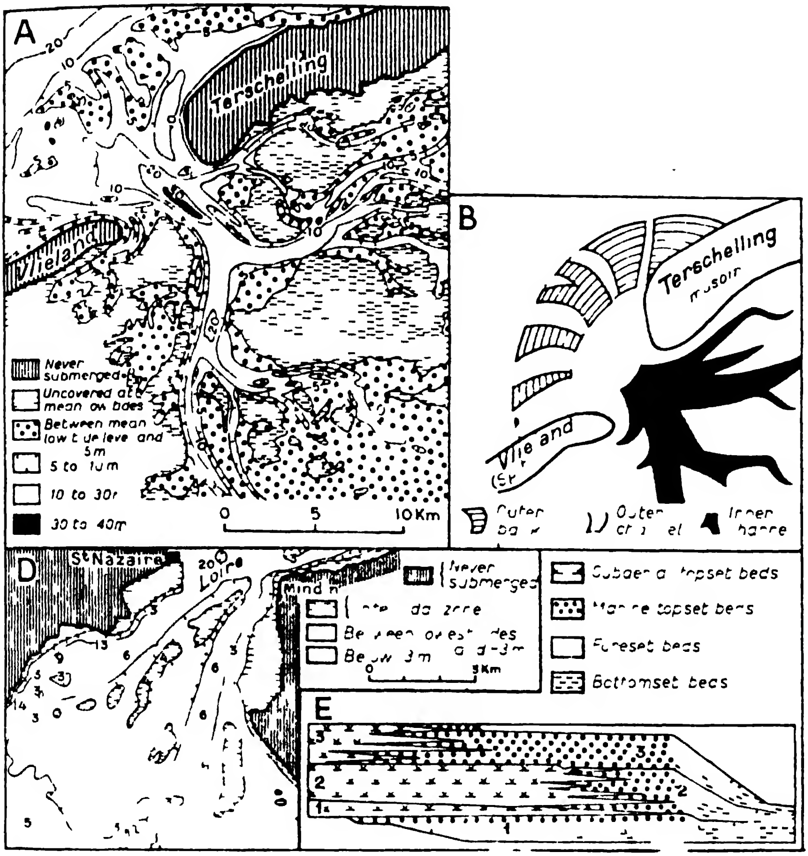


FIG. 13. SUBMERGED DELTAS: THE EFFECT OF SUBSIDENCE IN DELTAIC BEDS

A. Double tidal delta (internal and external) in Friesland (from Waddensymposium, *Tijd. Ned. Aard. Gen.*, 1950).

B. Diagrammatic representation of such a delta (after van Veen, 1936 and 1950).

D. Submarine delta of the Loire.

E. Relations between delta-building, marine transgression and subsidence (after Barrell (1912) in Twenhofel). 1. Delta built in relation to stable base-level. 2. Transgression or subsidence balanced by deposition. 3. Transgression or subsidence gaining over deposition with great extension landwards of marine topset beds.

Other submarine deltas occur in front of the Zealand estuaries. Associated with the passages between the Friesian islands, or with those through the offshore bars of the east coast of the United States, are submarine deltas of a different type, tidal deltas (Fig. 13 A). These are normally double, a larger delta in the lagoon and another smaller one outside separated by deep tidal channels, e.g. 47 m. in the Marsdiep in the north of Holland.

The submarine bar differs from the submarine delta in its form: it is more linear. A very characteristic bar is that of the Gironde: it also has two lateral channels, the channel at present in use through the middle being artificial. The bar is to be attributed to the piling up of sediments by westerly storms.

But, since these forms are below sea-level, they do not constitute real deltas. Even if the bar grew above sea-level and left a fairly wide entry and a large stretch of lagoon behind, it would not constitute a real delta: inlets of this type on the Black Sea coast are called *limans*. Real deltas lie above sea-level, or roughly at sea-level, as in the Danube.

C. K. Gilbert, in 1885, clearly described the method of formation of a delta, and it is worth while to return to the neglected pages of this great geomorphologist. The formation of a delta depends, he says, almost entirely on the following law: the capacity and competence of a stream vary with the velocity (the capacity is the total load of a given calibre which the stream can carry; the competence is the maximum size of the particles which the stream can move on the bottom). Thus, if the velocity diminishes quickly at the mouth, deposition will occur on the bottom of the channel. These deposits on the bottom form an obstacle to the current and promote continuous deposition upstream until the profile of the river acquires a continuous slope upstream from the mouth to some more steeply inclined part of the course. The slope must be sufficient for the velocity of the stream to be adequate for its load. Matters do not work out exactly like this but the stream constantly maintains a balance between its slope and the work it has to do. Since deposition starts at some distance from the mouth the decreased load may be transported over a gentler gradient: hence the profile becomes slightly concave. At the outer end of the deposit, there is often a sharp double break of slope: the slope of the surface of the stream changes to the horizontal surface of the sea or lake—any tide will therefore be a disturbing factor—and the slope of the stream bottom changes to the much steeper one of the delta front (see below).

As the current is swifter at the centre than at the sides, natural banks

or levées are built up on each side. However, these banks are never as high as the level of the greatest floods throughout. Therefore, accidental changes occur: the banks are breached, the breach is rapidly eroded, and the new channel is cut deeper than the old one which is quite likely to be abandoned. This is all the more likely if the old course has a very gentle slope, while the new channel presents a shorter and therefore steeper route to the sea. The position of the mouth varies and the stream tends to build up a semicircular delta, comparable with a waste fan in form but much flatter. In a river of considerable size, important displacements result in the creation of successive deltas. The Mississippi has six, of which the last but one is easily recognized to the north of the present one. The deltas are divided into sub-deltas and the present delta is a sub-delta of the last but one and has developed to a large size without further subdivision (Fig. 14 B, C and D). If not controlled, the river tends to flow towards the south-east through the Atchafalaya mouth. Old branches may continue to function to some extent, e.g. the Petit Rhône, the importance of which is, however, progressively decreasing.

In the huge deltas of the Siberian and American Arctic, alterations of channels are much more common than elsewhere, and are on a large scale. The climate and the régime of the rivers are the causes. At the time of the spring flood, when floating tree-trunks and ice erode the banks, enormous breaches lead to the cutting of new channels in the still frozen alluvium at the end of the spring. The pattern of land and water is extraordinarily complicated and the general outline can only be seen from the air.

On the levées and in the bed, the particles are quite large—sand and occasionally gravel. But beyond the levées are marshes and es, where the waters penetrate only after overflowing the levées during floods and into which only the finest sediments, i.e. mud and organic matter, are carried. As the branches are abandoned and cut off when their length becomes too great, deposits of former banks and old beds are found under lake deposits so that, in detail, the lithology varies abruptly. Periodical erosion and slumping of the banks create gullies and thus lead to false-bedding. Some delta lakes are very large, like the Vaccarès, or better still, Lake Pontchartrain on the Mississippi, which is 40 km. by 65 km. The largest and deepest tend to increase by erosion of the banks, i.e. where the fetch of the waves is long enough. The Vaccarès (Fig. 14 A) has progressively increased in size since the sixteenth century: it is bounded by minor, but nearly vertical cliffs where *Salicornia* and

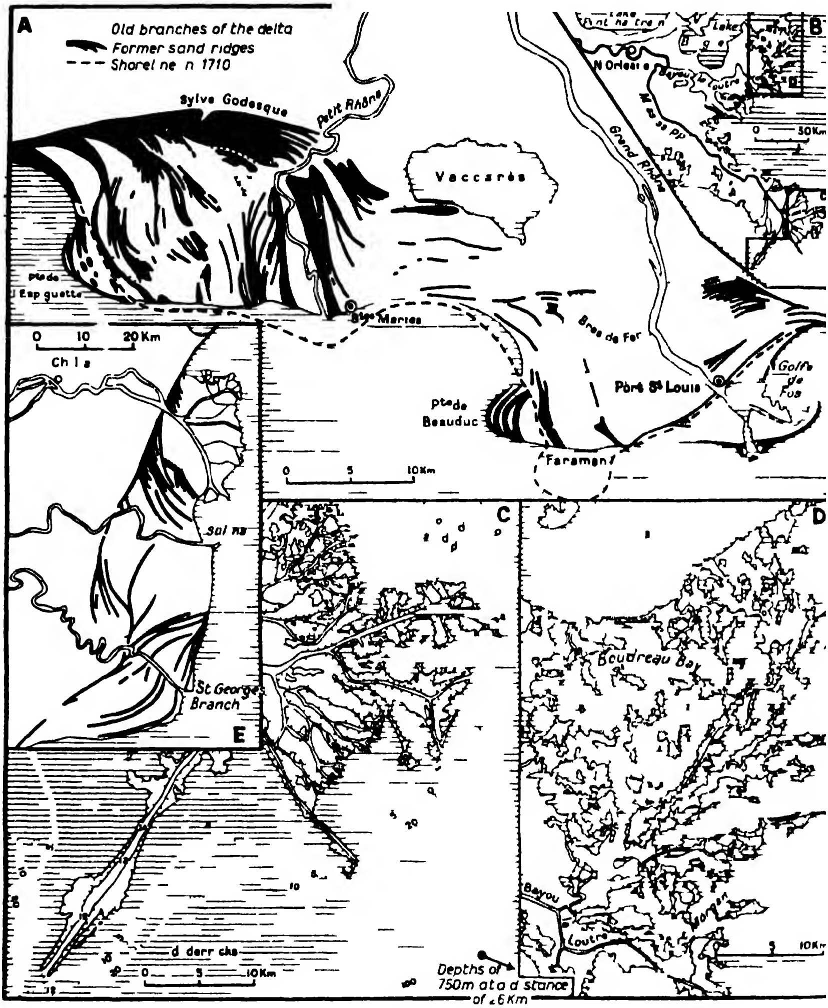


FIG 14 DELTAS

A Evolution of Rhône delta and former shingle ridges (after François, 1937 and Kruit, 1951) only the Vaccares lagoon is shown B General form of Mississippi delta C Modern part of the Mississippi delta D Old degraded part of the Mississippi delta positions of former branches are clearly shown E Danube delta (after de Martonne, 1931, Valsan, 1938, and Petrescu, 1948) former sand ridges are shown

tamarisks are undermined. This and the other lakes of the Camargue fill up as a result of organic sedimentation, aeolian deposits, and the waste eroded from the banks. In the Mississippi delta, similar phenomena involve the coalescence of originally separate lakes. In the smaller lakes the fetch is generally insufficient for marked erosion of the banks and they are more quickly filled by organic sediment because they are not so deep. Vegetable matter accumulates on the bottom and as the bottom rises, reeds appear in the swamp and finally a fen develops.

All these deposits form the topset beds. But there are also foreset beds and bottomset beds (Fig. 13 E). The foreset beds are steeply inclined, at angles of 10–25°, in the fossil delta of Lake Bonneville, where they are easy to study. Their slope is that of the angle of rest of the material concerned. Sediments are generally coarser in the foreset beds than in the topset beds except for the channel deposits. In the bottomset beds colloidal matter is dominant. These last are formed mainly in marine conditions, and partly owe their agglomeration, if not their deposition to flocculation by the salts in the sea-water. They are covered progressively by the foreset beds and these in their turn by the topset beds.

Mechanical action of the sea. But the sea or lake into which the river flows causes both erosion and redistribution of the material.

(*a*) Redistribution consists essentially in the building up of spits or bars which rapidly rise to the surface and cut-off lagoons which are finally completely separated from the sea: in this way lagoons of a type different from those between the branches of the delta are formed. Some are of mixed origin. The Rhône delta, especially in its north-west part, is quite remarkable in this respect (Fig. 14 A): aerial photographs show a large number of successive spits, the oldest of which is the Sylve Godesque near Château-Davignon. Fine modern spits are seen on both sides of the Mississippi: on the west, the Ship Shoal is a submerged offshore bar. The Danube delta is also built up of successive ridges known as '*grinds*', which form a goose-foot pattern between the arms (Fig. 14 E); each series of ridges was formed by the rearrangement of the deposits of one branch of the delta (Petrescu).

(*b*) Marine erosion working on a delta results in the wearing away of the topset and even the upper part of the foreset beds. Such erosion results from a general reduction of the deposits brought by the river, either through capture or a change of climate or, more often, through the abandonment of part of the delta after the river cuts a new channel.

Thus, the last but one delta of the Mississippi is being appreciably eroded (Fig. 14 D).

On the Rhône delta (Fig. 14 A) these processes are most characteristic. The erosion results from the following facts: abandonment of the Bras-de-Fer, or Vieux Rhône, in 1711; the decreasing importance of the Petit Rhône in relation to the Grand Rhône. On the coast itself there are two important winds; the mistral, a powerful north-west wind with only a short fetch, and the marin, an onshore south-east wind which is less common but has a much longer fetch. The chief swell therefore comes from the south-east.

When the Grand Rhône flowed out by the Bras-de-Fer, the Faraman side had a very pronounced point, more than 5 km. in front of the present coast. The point in the Gulf of Fos did not exist at that time. The points of Beauduc and l'Espiguette hardly existed and the Petit Rhône, still active, had built up a headland some 2,100 m. in front of its present mouth.

Since 1711 the points of the Bras-de-Fer and the Petit Rhône have been worn back owing to an absence of sedimentation and the deposits thus set in motion, together with those from the new mouth of the Grand Rhône, have given rise to the building up of the points of Beauduc and l'Espiguette. This evolution has resulted from the action of the south-east swell which causes, by its oblique approach, a drift of material towards the west. The Faraman lighthouse has had to be moved 1,200 m. inland. But when it reaches the bend of the coast at Beauduc, the drift is stopped by the secondary swell from the north-west, caused by the mistral in the Gulf of Beauduc. Thus a headland is formed and gradually built out into the Gulf of Beauduc. Only its end advances, since the coast further to the north is eroded by the north-west swell. L'Espiguette evolves in exactly the same way. The general evolution is like that at Dungeness, and clearly shows the predominance of marine action. All the river does is to vary the locality where alluvium is brought into the sea.

The rate of marine erosion in deltas varies with the material attacked. In sand and gravel it is quick. In clays, which may be extremely tenacious, it is clearly slower.

Classification of deltas. Based on the relative effects of the rivers and the sea, and the differences they produce, Gulliver put forward in 1899 a classification of deltas which is reproduced here with slight modification.

(*a*) Deltas where river alluvium is abundant are lobate: either with

VIIA Underwater overhang on coral reef on the Farsan Bank, Red Sea

VIIB Coral reef on the Farsan Bank at a depth of 2–3 m
(Underwater photos Nicolleau

one lobe like the Meander, or with many lobes like the Mississippi, the Po, the Rhine, and the Volga. In extreme cases the banks enclosing the branches form long fingers into the sea, as in the Grand Rhône and the Mississippi.

(*b*) When marine erosion is more important, the convex lobes are replaced by concave crescents facing the sea and often meeting at the mouths of the streams, as in the Tiber. The Nile and Tagliamento show features of both type (*a*) and this cuspate type.

(*c*) With active coastal currents, rounded deltas are formed, e.g. the Arno, Llobregat, Rion, and Crati.

(*d*) Almost complete dominance of marine action only allows a stunted delta to form, e.g. the Garigliano, Guadalaviar, and Têt.

Where marine action is absolutely dominant, the mouth is completely barred and the river escapes by seepage through the bar. This case, already considered in the section on beaches, is no longer a delta.

Relative variations of the sea-level and subsidence. When they begin to form, deltas soon develop above sea-level but the submerged part only lengthens and thickens later. If the land undergoes a slow and fairly prolonged uplift, the emerged part of the delta may be readily built up provided that the supply of material is sufficient.

If the sea-level rises as the delta is forming, the underwater part thickens. During intermittent transgressions, more rapid than the rate of upgrowth of the delta, the sea advances and causes erosion on the top of the delta especially on its outer edge and also some deposition of marine sediments on the eroded surface of the delta. When the transgression stops, the delta grows forward again. Such successions are found in the Eocene in the Hampshire basin, where marine beds thinning westwards alternate with deltaic beds thinning eastwards (Stamp, in Twenhofel).

In certain large deltas continuous subsidence allows the accumulation of considerable thicknesses of material. In the Mississippi delta, a thickness of 900 m. is attributed to recent deposits, 900 to the Low Terrace, and 700 to the one preceding. The base of the Cretaceous is at a depth of more than 8,000 m. Such great accumulation is certainly due to a subsidence which may be explained, but perhaps only partly, by isostatic compensation (see Chapter VII) caused by overloading. The result is that the delta remains permanently in approximately the same place.

D. CORAL FORMATIONS

Coral structures form a distinct class of coastal phenomena since they are mostly of organic origin and are not principally dependent on the action of the ordinary coastal processes: this, however, does not mean that the latter play no part in their formation.

Suitable conditions for coral growth: Corals build reefs only in limited areas, determined by the temperature, depth, movement, salinity, and turbidity of the water. Outside these limits isolated corals exist in certain places but are of no importance as reef forms. Attention must also be paid to the calcareous alga, *Tenarea tortuosa*, of the Mediterranean, which builds a ledge from 0·50 m. to 2 m. wide (p. 66).

Coral reefs, which are known from strata as old as the Cambrian, develop in seas where the temperature never falls below 18° C., and remains on the average a few degrees above that figure. The most favourable temperature is from 25 to 30° C. Light is essential since the polyps live in symbiosis with unicellular algae, the *Zooxanthellae*, which need strong light for their functions. This is why the chief development of reefs takes place between ordinary low tide and a depth of 15 fathoms. At this depth, the number of surface reef species falls abruptly, and there progressively appear non-surface species which live under different conditions; the reef-building corals are practically restricted to comparatively shallow water. Above ordinary low water level corals cannot live, since they cannot bear any prolonged emersion, and temperatures of more than 36° C. are fatal.

Corals like disturbed water that is continually being changed, since it contains more oxygen and is richer in nutritive matter; but breakers which are too strong are unfavourable since they may damage the coral. Salinity must be between 27 and 40 per mille. Turbidity is generally unfavourable, because coral dies if choked with mud. However, different corals react differently. Some are almost incapable of combating silt; others, like *Fungia*, combat it effectively by the movement of their cilia. Mud falling from above may be eliminated in this way; but corals can do nothing against mud which buries the base of the organism. On the whole much mud or sand is unfavourable, but coral reefs do exist in the proximity of certain river mouths as in Indonesia and Hawaii, and in front of the River Loza in Madagascar.

It has often been said that coral requires a solid base for its growth. This is true where the waves are powerful enough to destroy reefs

which are not fully established, but in the calm water of the coral lagoons of the Red Sea, it is not unusual to find small colonies merely resting on sand so that they can be easily moved by hand. In Australia borings in the Great Barrier Reefs show intercalations of sand and mud between beds of reef corals and in Indonesia certain reefs also rest on silt or mud.

The world distribution of reefs depends upon these requirements: coral reefs are found in tropical seas from Bermuda, the north of the Red Sea, Midway, and the Hawaiian Islands in the Northern Hemisphere, to Houtman's Abrolhos on the west coast of Australia and Lord Howe Island at 31° 30′ in the South Pacific between Australia and New Zealand. They are absent or rare on the eastern sides of oceans, as the currents there are usually too cool: for example, on the American Pacific coast colonies occur only between the Gulf of California and latitude 2° South. They shun the great deltas of India, Indochina, and China and those of the tropical Atlantic, they avoid confined coastal lagoons where the water is insufficiently aerated, too hot or too variable in temperature and salinity. They flourish, on the other hand, in the central and western Pacific, on the north-east, north and north-west coasts of Australia, in Indonesia, the islands and the west coast of the Indian Ocean, in the Red Sea, Eastern Brazil, and in the Antilles. In these places superb developments of coral islands and reefs are found, lagoons of an indescribable greenish blue, hemmed in by majestic and powerful breakers in the open ocean, gardens of stone where yellow, violet, green, and red intermingle. Massive and branching corals are intermingled and innumerable fish rival the coral in their magnificence.

Coral is always accompanied by numerous calcareous algae, the most important of which is *Lithothamnion*, and by gastropods, lamellibranchs, echinoderms, and foraminifera, which play a very large part in the construction and nature of the reef. The living part of the reef forms the *bioherm*, but a far larger part of it consists of detrital material derived from the bioherm by sea action.

Reef forms

1. *Atolls* (Fig. 15 A). Atolls are rings of coral interrupted by channels, surrounding a lagoon whose depth generally exceeds 30 m but rarely 100 m, and whose diameter, although very variable, may exceed 60 km. The channels are very rarely more than 100 m in depth. The true atolls are nearly all in the Indian and Pacific oceans and the Indonesian seas, however, the Red Sea has some typical examples such as Sanganeb in

front of Port Sudan. The zones on an atoll are usually as follows (Fig. 18 A):

(*a*) *The seaward slope*. This slope usually falls very steeply to considerable depths: it reaches and frequently exceeds an angle of 45° for several hundreds of metres, and even in the upper parts which can be explored in a diving suit, overhangs of several metres are frequently observed (pl. VII A). Overhangs are quite normal where the level of the sea remains constant. Corals in their struggle for light, oxygen, and nourishment must grow outwards and thus form an overhang, which may be concealed by detritus only. The detritus is built into the reef by the action of calcareous algae, *Alcyonaria*, etc., so that the outside slope becomes compact and jagged apart from the sand which covers ledges occurring on the slope.

(*b*) *The algal ridge*. This ridge, which is clearly marked only on the windward side of the reefs, is also called the *Lithothamnion* ridge, but actually consists mainly of *Porolithon*, another calcareous alga. It forms on the outer edge of the reef. It is the highest part of the atoll apart from any true island included in it. It is absent from the Indonesian and Red Sea reefs, as it is an oceanic form, formed in the zone where long, powerful swell breaks, and such swell is absent from enclosed seas. The surf is very powerful since there is no platform in front to weaken the swell. The surf attains a height of several metres, and makes the reefs inaccessible from windward, except through the channels, and considerably disturbs coral growth down to depths of 2 or 3 m. But *Porolithon* flourishes in these breakers. It can live, unlike coral, above sea-level so long as it is constantly moistened by spray. It therefore forms a ridge on the windward side, as at Bikini in the Marshall Islands, where it builds a red or rose-coloured rampart, like a cuesta in form, facing the open sea, and lying 0·6–1 m. above the reef-flat behind it.

At Bikini the top of the outside slope on the windward side is intersected by regularly-spaced submarine grooves and intervening spurs. These seem to be an adaptation to the breakers, which they canalize, and cause to break early. The grooves extend inside the algal ridge as surge channels, not as a result of erosion (according to Emery, Tracey, and Ladd), but because *Porolithon* forms bosses and ridges between them and gradually grows over them, thus forming in places what is called a 'room-and-pillar structure'. The water brought by the waves spurts up like geysers from the inner ends of these tunnels, around which calcareous algae construct raised rims on some reefs; the water runs away into a succession of basins bounded by walls deposited by the

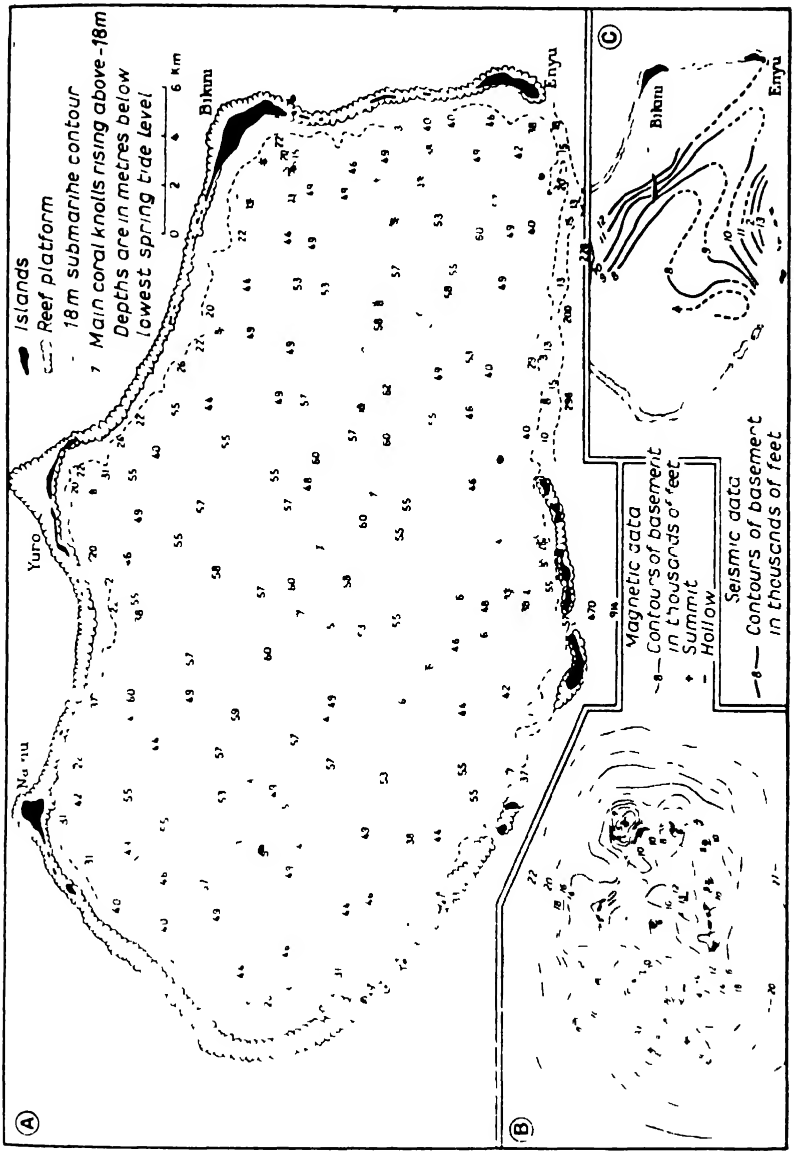


FIG. 15 BIKINI ATOLL

A. General map (after U.S. Hydrographic Office). B. Magnetic results of Alldredge and Dichtel (1949). C. Seismic results of Dobrin, Perkins, and Snavely (1949). The contours of the basement obtained by these methods are approximate.

same alga. On the leeward side of the reef, the algal ridge is much lower, and the grooves are absent or replaced by caverns enclosed by coral and calcareous algae. These forms are not peculiar to the Marshall Islands, but also occur, for example, at Funafuti in the Ellice Islands, at Raroia in the Tuamotu group, at Onotoa in the Gilbert Islands. However, there is much controversy among scientists whether they are mainly constructional or erosional forms. In any case, exposure often plays an essential part in the formation of reefs.

(*c*) *The reef-flat* lies behind the algal ridge and consists of different subzones which vary with locality. It is usually several hundred metres wide. This platform, which is very irregular and uneven in detail, is partly visible at low water, especially in places, such as parts of Australia, where the tidal range is fairly large. It is formed mainly of dead coral—bored into mechanically or chemically by algae, sponges, and *Lithophagus*—and also of detritus re-cemented and encrusted by calcareous algae. The whole platform is often partly occupied by living corals, which near the algal ridge find favourable conditions for life in the abundant supply of water from the ocean. This coral, when it consists of *Porites*, often assumes the shape of 'micro-atolls', circular constructions of coral, in which the live individuals are at the edge, which thus rises a little above the centre. The height is usually a few decimetres and the diameter varies from several decimetres to several metres. The hollows are partly filled with calcareous sand. Accumulations of detrital blocks form banks in places. The largest, broken off by hurricanes or typhoons, are the 'negro-heads'. Some negro-heads, however, which form an integral part of a reef, are in fact all that remains of an older reef, built in relation to a higher sea-level, and now reduced to present sea-level as a result of corrosion after a slight fall in the level of the sea.

On reef platforms islands are found. They are formed of calcareous sand produced by the wear and tear of corals, foraminifera, and algae. The sand on the beaches is often consolidated into a calcareous sandstone, the beach-rock, which is still forming, but is often cut into jagged forms by corrosion (see p. 29). Islands may shift in the direction of dominant winds, and all that remains to indicate their former positions are one or more ridges of beach-rock.

(*d*) *The inner slope* is more gentle than the outer slope. The upper part is formed of sand swept from the islands. On the slope there are colonies of living coral with large flat-topped heads which almost reach the surface. The appearance of the slope and the creatures living there

vary, depending upon whether they are on the windward side, where they are under the shelter of the reef, or subject to the waves formed in the lagoons.

(*e*) *The lagoon* has variable relief. The bottom is sometimes flat and covered with calcareous sand of unknown thickness, constantly triturated by *Holothurians* whose morphological effect, however, cannot be very great. But the sandy bottom is often broken by numbers of steep knolls of living coral which rise to within a few fathoms of the surface. The Eniwetok lagoon in the Marshall Islands, of which Emery has produced a very fine chart, contains about 2,300 of them. We must not therefore assume that coral does not grow freely in lagoons.

2. *Barrier reefs*. These reefs enclose one or more non-coral islands. In detail, the zones and general form are much like those of the atolls. The Queensland Barrier Reefs, the most famous in the world, consist of two very different sections; the first north of Trinity Opening and the second south of it and including the Capricorn channel. Corals do not form true barrier reefs on the other coasts of Australia. Other fine barrier reefs encircle New Caledonia, and the Fiji Islands. Barrier reefs around very small islands closely resemble atolls, for example in the Truk Islands in the Caroline group. Double barrier reefs, like that off the north-west of Viti Levu, Fiji, are extremely rare.

3. *Fringing reefs*, which lie close to the shore of the land, are one of the commonest forms. Two types may be distinguished.

(*a*) *Reefs facing the open sea and unprotected by a barrier* show a series of zones analogous with those of barrier reefs and atolls.

(*b*) *Reefs protected by a barrier* are extremely common. Sheltered from the great breakers, they have no algal ridge, but their outer edge is often very abrupt. They resemble the reef-flats of atolls in being very irregular, and are occupied by a series of pools at low water.

Very often the fringing reef is not actually joined to the coast, but separated from it by a channel 0·3–1·5 m deep, called the 'boat channel', the bottom of which is covered with sand and gravel, and sometimes plants, e.g. in Madagascar. This is in many cases connected with the fact that the abundance of sediment near the shore is unfavourable for the growth of coral. The east coast of the Red Sea north of Jiddah has a very long channel of this type (Fig. 18 B), which occurs also in the fringing reefs of the north-west coast of Madagascar.

On any one island, a reef may be successively a fringing reef, and a barrier reef, and may even simulate an atoll at some point: examples occur in the Palaos Islands, Wallis Islands, and the Fiji Islands (Fig. 17 C).

4. *Shallow lagoon reefs*, which are annular, are like small atolls in which the lagoon is only a large pool a few decimetres or metres in depth. This type of reef is almost absent from the open oceans although some of the atolls of the south-east Gilbert Islands are more or less of this nature. By contrast, such reefs are very common in epi-continental seas such as the seas around Indonesia, the South China Sea, in the Antilles, on the north part of the Farsan bank in the Red Sea (pl. VIII A), along the north-west coast of Madagascar, and especially round Australia, where they have been carefully studied, notably between the Great Barrier of Queensland and the mainland. These reefs, like real atolls, are found in situations exposed to wind and swell. Off Queensland, they are in the path of the south-east trades. They are usually oval, with the major axis orientated north-west–south-east, and are higher on the windward side. From south-east to north-west the following zones are found (Fig. 16 A).

(*a*) Above the outside slope, which is covered with living corals and rises rapidly from the relatively shallow depths of a few dozen metres found in the steamer channel, concentric ramparts of coral debris have been built up one after the other. Some vegetation, generally *Avicennia*, colonizes the ramparts, which die out towards the north-west. The oldest are cemented into a conglomerate.

(*b*) Behind the ridges there is a pseudo-lagoon or swamp, the bottom of which is covered with black calcareous mud and sand. At the south-east corner of the swamp grows a high, dense cover of mangrove with *Rhizophora mucronata*. This pseudo-lagoon partially dries out at low water.

(*c*) At the north-west or leeward end there is a sand cay which may or may not be wooded, but usually includes beds of beach-rock. According to Steers, the cays are formed where the sand is prevented from migrating farther to the north-west by the swell refracted or diffracted in the lee of the reef. The cay is absent from poorly developed young reefs. The whole reef is of moderate size, its longer axis being only a few hundreds or thousands of metres long.

These forms seem to be associated with constant or prevailing winds, for example, on the northern part of the Farsan bank in the Red Sea, where the prevailing wind blows from the north-west, and where contrasts between windward and leeward slopes, similar to those at Bikini, are found (Fig. 16 C). The differences between these and the Queensland coast forms are not fundamental: in the Red Sea the underlying solid rock is at a greater depth, mangroves are absent, and the

VIIIA Marmer Reef, Farsan Bank, Red Sea. The white area is a shallow lagoon and the black area the central cay

VIIIB Contraposed coast at Porth Neigwl, Lleyn peninsula, Wales. The bay is cut into glacial deposits, subject to slipping, and the headland is formed by hard rock. (*Photos: A. Guilcher*)

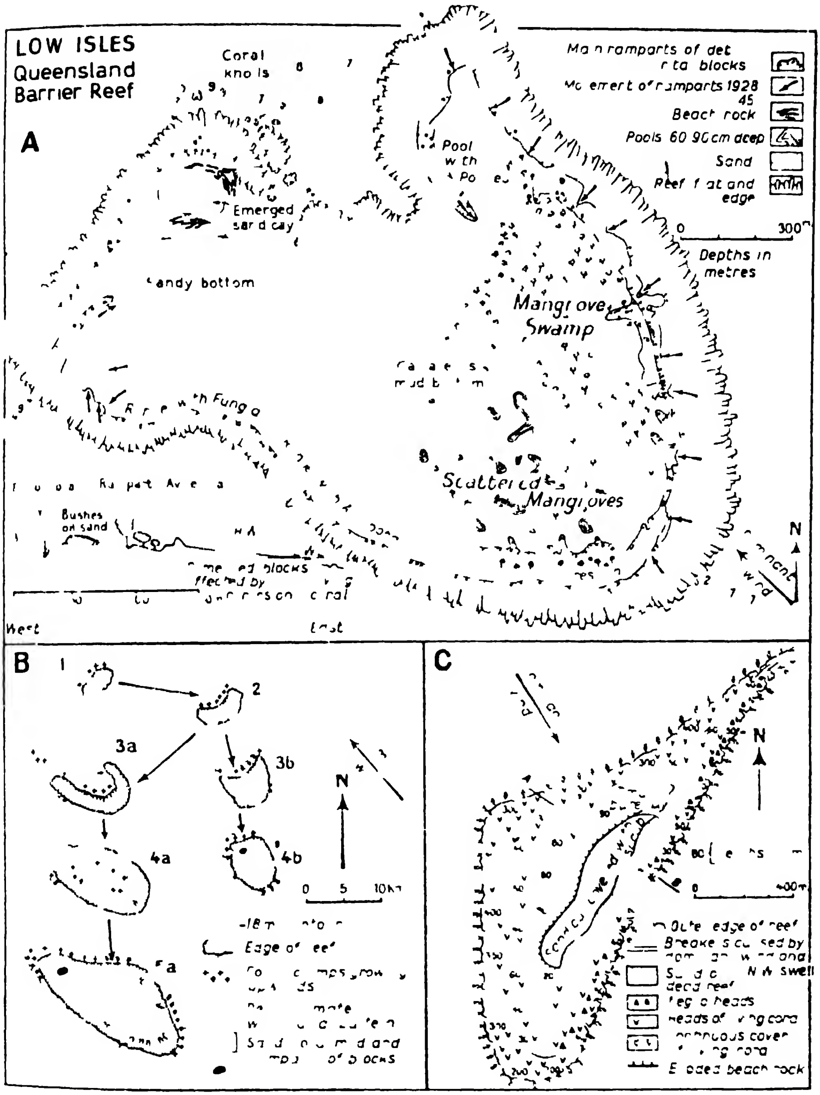


FIG. 16 ANNULAR REEFS WITH SHALLOW LAGOONS

A Low Isles, Queensland Barrier Reefs (16 23 S, 145 34′ E) (after Fairbridge and Teichert, 1948) B Mode of evolution of such reefs (after Fairbridge, 1950) C. Marmar reef at the northern end of the Farsan bank in the Red Sea (after Guilcher, 195

banks of shingle are obscure and below sea-level. In Indonesia and the South China Sea climatic conditions are different: as a result of the monsoonal régime of seasonally changing winds, cays which are not fixed suffer a seasonal change in orientation. If the two monsoons are of equal strength, the cay is central, and there are ridges of debris on both sides of the reef. The form is then symmetrical as in the South China Sea. If the monsoons are of unequal strength the forms approach those of Queensland or the Red Sea. Verstappen has shown that the Indonesian reefs of this kind may be greatly modified in shape if the prevailing monsoon changes for a number of years.

Small reefs of the same kind may be associated with strong tidal currents, rather than with the wind; for example certain reefs between Cape Direction and Cape Melville on the Queensland coast, where structural control may also play a part, and certain Indonesian reefs such as those in the Sibitu archipelago to the north of Borneo. In these conditions they form either annular reefs extending in the direction of the current and separated by narrow, deep channels sometimes exceeding 100 m. in depth, or simple elongated reefs of the same type.

5. *Faros* are chains of small atolls with lagoons of no great depth, forming either a large atoll or a barrier reef. This curious type of reef is not very common. It is found in the Moluccas, e.g. Maria Reigersbergen in the Banggai archipelago, and on the Tagula barrier to the south-east of New Guinea, and the Vanua Levu barrier in the Fiji Islands. The most numerous and complex examples are in the Maldive archipelago, from which the name faro is derived. The southern Maldives have no clear faros, but north of the Veimandu channel they include (Fig. 17 A and B):

(*a*) Two large circular groups of atolls stretching 260–300 km. north–south and about 90–100 km. east–west.

(*b*) Each of these two groups contains seven or eight large atolls, the lagoons of which are generally 35–70 m. deep; between these atolls are channels generally more than 400 m. deep. Comparable depths occur in the centres of the two large circular groups.

(*c*) Faros occur on the edges of the large atolls. Some are 10 km. or more in length, but the majority are smaller. The smallest are almost circular, but others are elongated parallel to the edges of the larger atolls. Usually their lagoons are less than 20 m. deep, but vary between 5–6 m. and almost 40 m. Shallow lagoons are also characteristic of faros in other districts and to some extent they may resemble the annular reefs of Australia. In addition, several of the larger atolls in the Maldives,

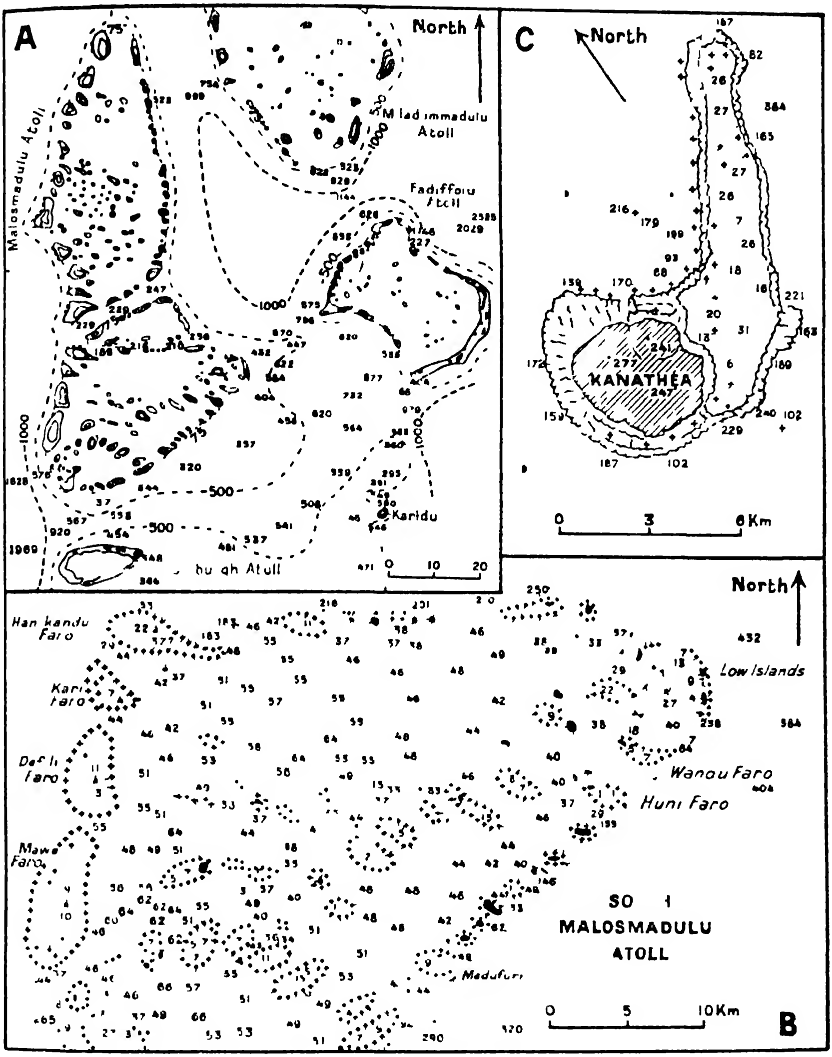


FIG. 17 FAROS, BARRIER REEFS, AND FRINGING REEFS

A. Northern Maldives (after Kuenen, 1950)—based on Admiralty chart No. 66A. Depths in metres. B. Detail of one atoll and associated faros in same group: based on Admiralty chart No. 66A. Depths in metres. C. Kanathea, Exploring Islands, Fiji group (17° 1? S., 179 10′ W.)—based on Admiralty chart No. .6. Barrier and fringing reefs, passing into an atoll form in the north-east. Steep outer slopes. Depths and heights in metres.

notably South Malosmadulu, Ari, and Nilandu, contain many faros in their lagoons beside those occurring on their outer edge. The channels between the marginal faros are 25–60 m. deep, i.e. about the same or a little less deep than the lagoons of the large atolls.

6. *Coral banks* are isolated, rather shapeless reefs. They may be embryo forms of the shallow lagoon reef, as often appears to be the case in Australia, or atolls with filled lagoons as in the south-east Gilbert Islands, or parts of fringing or barrier reefs isolated from the main structure, or the upper parts of coral knolls in lagoons.

7. *Submerged and raised reefs* are old reefs, no longer being built up because conditions under which coral can live are no longer present. In this class are forms such as the drowned atolls of the west central Carolines where one of them covers 885 square miles, and the drowned barrier reefs like that of Waigeo in the Moluccas where the difference in depth between the lagoon and the enclosing reef is 170 m. in places. The submergence may be due to a subsidence or a marine transgression too rapid for coral to continue growing. Raised reefs are very numerous even if we exclude fossil forms, in which large colonies may affect the course of erosion. Examples of recently uplifted reefs are the barrier reefs of New Georgia in the Solomon Islands, and of Mangaïa in the Cook Islands; the Guam reef which is tilted; the Poeloe Dana reef west of Timor which is an atoll with a dried-up lagoon; the Abulat reef on the Farsan bank in the Red Sea which is broken by numerous faults into small tilted blocks. When they are well preserved they retain the steepness and the overhanging sections of the former submarine slopes and sometimes form high steep cliffs, as at Makatea in the Tuamotu group.

The origins of reefs. The problem of the origin of reefs has already been touched upon; it has been seen that the form of reefs is influenced both by swell and at times by currents. These factors satisfactorily explain the form of annular reefs with a shallow lagoon, while, in Australia, their growth from small coral flats has been traced (Fig. 16 B). Can we also attribute atolls, which are also controlled by the swell, to a similar process of the merging of the ends of an original crescentic form open downwind? Certain writers such as Krempf and Wood-Jones have thought so. But the great difficulty is that atolls are evidently akin to barrier reefs, for which the foregoing explanation is of no value. It does not explain the distance between the central island and the barrier reef on the windward side. We must also assume that reefs in the open ocean rest on rocks other than coral even when the underlying

rocks are not visible. We also need to know why the underlying rocks are not visible and at what depth they are to be found.

The first serious explanation was Darwin's. It may be summed up in one word: subsidence.[1] The coral first of all grew as a fringing reef round an island. Then the island subsided and the coral grew vertically upwards or even somewhat overhanging on the seaward side. Thus a barrier reef was formed. If the subsidence was sufficient for the island to disappear an atoll was formed.[2] Although at first received with admiration, Darwin's explanation, which was adopted by Dana, was subject to considerable criticism towards the end of the nineteenth century, and by 1900 was scarcely believed. Yet, as Davis pointed out, when he revived the theory, Darwin had foreseen all the objections which were made to his views.

The theories proposed at the end of the nineteenth and the beginning of the twentieth century to replace Darwin's have been summarized by Davis in his main work (p. 22). Reefs may have been formed

'by upgrowth on still-standing, aggraded submarine banks, as Rein first and Murray later suggested and as Wood-Jones still more recently argued; by outgrowth from still-standing foundations, the lagoon being excavated by solution back of the living reef, as Murray also proposed; by outgrowth on rising foundations, as Semper and Guppy believed, Semper having anticipated Murray in explaining lagoons by solution; by upgrowth on still-standing foundations completely truncated by abrasion, as Wharton proposed; or incompletely truncated as Tyerman and Bennet long ago and as Guppy later suggested; or on platforms produced around still-standing foundations by subaerial erosion and submarine denudation, as Agassiz believed . . . for all these theories succeed in explaining the visible features of reefs that they were

[1] Or more exactly: reefs growing on rock foundations subject to intermittent or continuous subsidence.

[2] Davis (1928, pp. 131–7) maintained that if subsidence is continuous and slow enough the reef grows first outwards then upwards and finally inwards, so that in cross-section a convex slope is formed. First of all (Fig. 18 C), the amount of material required for the building of the external talus slope (and also the filling up of the lagoon) is not large, but it increases progressively as the reef grows up. The convexity of the reef profile results from the fact that, as the surface area of the talus slope increases, the same amounts of debris form progressively thinner layers on that slope. But this deductive argument, although logical, is entirely hypothetical, and is not in agreement with all the observed data. In fact, the common occurrence of steep outer slopes, often hundreds of metres high, suggests that the supply of talus plays a less important role than is implied by Davis, and that vertical growth continues for a longer time than Davis thought likely. Consequently Fig. 18 C is a modification of Davis's figure.

invented to explain. It is with regard to the invisible conditions and processes of the past that the theories differ.'

Davis has shown that Darwin admitted the possibility of the first theory applying to certain cases. It was later revived by Hoffmeister and Ladd in the 'Theory of antecedent platforms', which covers all reefs establishing themselves on any foundation situated at a suitable depth and so offering favourable conditions for coral growth. The idea that the lagoons are due to solution has also been upheld by Stanley Gardiner, but it creates difficulties if it is combined with the first theory quoted by Davis above, for (Davis, p. 60) 'the same solvent action ought to have prevented the upgrowth of the bank [by submarine sedimentation] before it reached so small a depth that corals would settle on it'. Further, recent work in the Pacific, notably at Eniwetok, shows clearly that lagoons are being filled up by the growth of coral and not being enlarged. The third theory, which has been applied to the Palaos and Solomon Islands, has been rejected for those places by Davis on the grounds that the volcanic substratum of the raised coral reefs of these islands suffered subaerial erosion before the corals gained a footing there. These were, without doubt, subaerial volcanoes; later these volcanoes may have subsided, thus allowing coral to grow there (according to Darwin's theory), and finally the volcanoes together with the reefs have been uplifted.

The glacio-eustatic theory. A more recent theory and one which needs close examination was put forward by Daly, who explained most of the characteristics of reefs by glacio-eustatism. His explanation is based upon the fact that a great number of lagoons are about 60 m. deep. These are thought to be non-coral platforms merely covered with a thin layer of debris and calcareous algae. They were eroded during a glaciation, when sea-level fell, in marginal belts of the coral seas in which the cold would have prevented corals from living. Afterwards the postglacial amelioration of climate allowed reef building to take place more or less at the same time as the sea-level rose. The corals are thought to have colonized the marginal regions from the warmer seas where they would have existed throughout. Daly calls attention to the widths of the atolls and barrier reefs, which are greater in the equatorial than in the marginal regions, thus implying that in the latter, the reefs are more recent. He admits earth movements but considers them to be only of local importance.

Without entering into the details of the discussion, which would be

impossible here, it must be stressed that the swing of opinion in favour of Darwin is due to Davis who, after an extensive voyage in the Pacific, and thirty-five articles dealing with very different coral regions, came down emphatically in his final synthesis in 1928 in favour of the theory of subsidence. He does not deny that the glacio-eustatic theory has its value, and thinks that it can explain the abrasion, at times of low sea-level, of the islands of the marginal coral regions. In these regions, the existence of cliffs cut into the islands behind the actual barrier reef, is, he says (p. 219), evidence in favour of Daly, for the reefs could not have been there when the cliffs were formed. But he thinks that abrasion during low glacial sea-levels was not very important. He questions, as others have done after him, the uniformity of depth of lagoons. He constantly uses an important argument in favour of Darwin's theory, an argument not put forward by Darwin but envisaged by Dana, namely the presence of embayed shorelines, i.e. submerged valleys, in islands surrounded by coral: entrenched valleys continued in long, deep bays prove submergence. After examining the possible effects of the post-glacial rise of sea-level, Davis concluded that this is insufficient to explain the features of submergence shown by islands surrounded by barrier reefs, and that subsidence must also be invoked to explain the relief.

Borings. Borings in reefs ought to be a decisive test of the theories invoking subsidence (Darwin, Dana, Davis) and of the glacio-eustatic theory (Daly). In fact, the latter implies that coral formations should not be thicker than the depth of water returned to the oceans by the melting of the glaciers, as corals are presumed to have formed only a thin cover before glaciation. Subsidence on the other hand implies a considerable thickness of coral, which, according to Daly, would be exceptional.

Borings, which are now fairly numerous in different regions, clearly favour subsidence. With the possible exception of the Bermudas which are only partly coral, few coral areas show features entirely in agreement with glacio-eustatism. The Bermudas stand on a platform at a depth of 75 m. and seismic work in 1952 showed a planation at this level under the whole of the archipelago, as Daly's theory implies. The boring of Kita Daito Zima, or North Borodino, south of Japan, shows Plio-Pleistocene coral down to 103 m., that is, at somewhat too great a depth to be explained by eustatism as Daly conceived it. This boring went down to 432 m. without reaching the basement. Two bores put down in the lagoon of the Great Barrier Reefs off Queensland, at

Michaelmas Cay and Heron Island, proved recent coral down to – 123 and – 145 m. At Funafuti in the Ellice Islands there is coral *in situ* down to – 339 m. at which level the boring stopped. At Maratoea to the north-east of Borneo coral is found as far as the bottom of the bore at – 429 m.; at Oahu in the Hawaiian Islands, down to – 319 m.; at Bikini in the Marshall Islands, the deepest of the four borings reached – 777 m. without entering the basement, and coral is reported from depths as great as – 640 m. (see *Bikini and nearby atolls*, pp. 82–3) Daly has suggested that some of these could be explained by assuming great banks of talus on the outside slope, but the Queensland borings are in the lagoon and not on the barrier proper; the Maratoea boring is in the lagoon of an atoll; at Funafuti, where the boring was started on the edge of the atoll, the coral is described as being *in situ* and not as talus, and at Bikini there is a considerable thickness of pre-Quaternary coral construction: the top of the Upper Miocene is tentatively placed at – 259 m., and the coral lying at – 622 m. is assumed to be Oligocene. Moreover, a magnetometer survey at Bikini shows that the foundation of this atoll, presumably volcanic, varies in depth from – 1,520 to – 3,950 m. Lastly, at Eniwetok in the Marshall Islands, two borings in the north-west and south-east of the atoll reached the foundation at 1,401 m. and 1,267 m. In the second, the foundation is olivine basalt, lying under shallow-water Eocene limestone. In both, clay, mud, and hundreds of feet of Quaternary coral limestone are passed through before the Tertiary, mainly limestone with a little dolomite, is reached. The Pliocene and Quaternary subsidence of Bikini and Eniwetok is therefore only a continuation of an earlier and very much greater subsidence (Fig. 15 B and C).

Borings prove, therefore, that the glacio-eustatic theory is insufficient and that one must revert to Darwin's theory in the majority of cases. In the case of Eniwetok the theory of antecedent platforms may also be applied. Moreover, it is difficult to explain in any other way the steepness of many of the outer slopes often hundreds of metres high, as these are steeper than submarine volcanoes and talus slopes.

Theories involving earth movements and eustatic movements. However, eustatic oscillations of sea-level in the Quaternary must have taken place and had some effect on coral formation, as Davis himself clearly recognized. Taking this into consideration, Kuenen and Stearns have each made attempts to develop new theories involving both isostatic movements and all Quaternary eustatic movements and not merely the last. Earth movements are of various kinds: according to Stearns

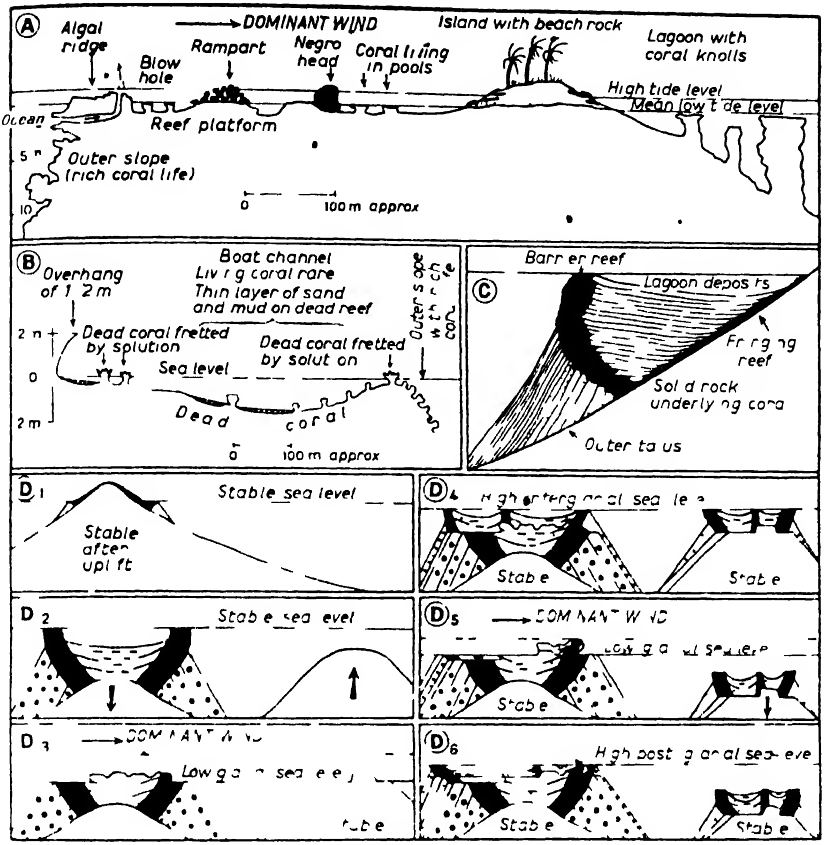


FIG 18 SECTIONS THROUGH REEFS THEORIES OF FORMATION

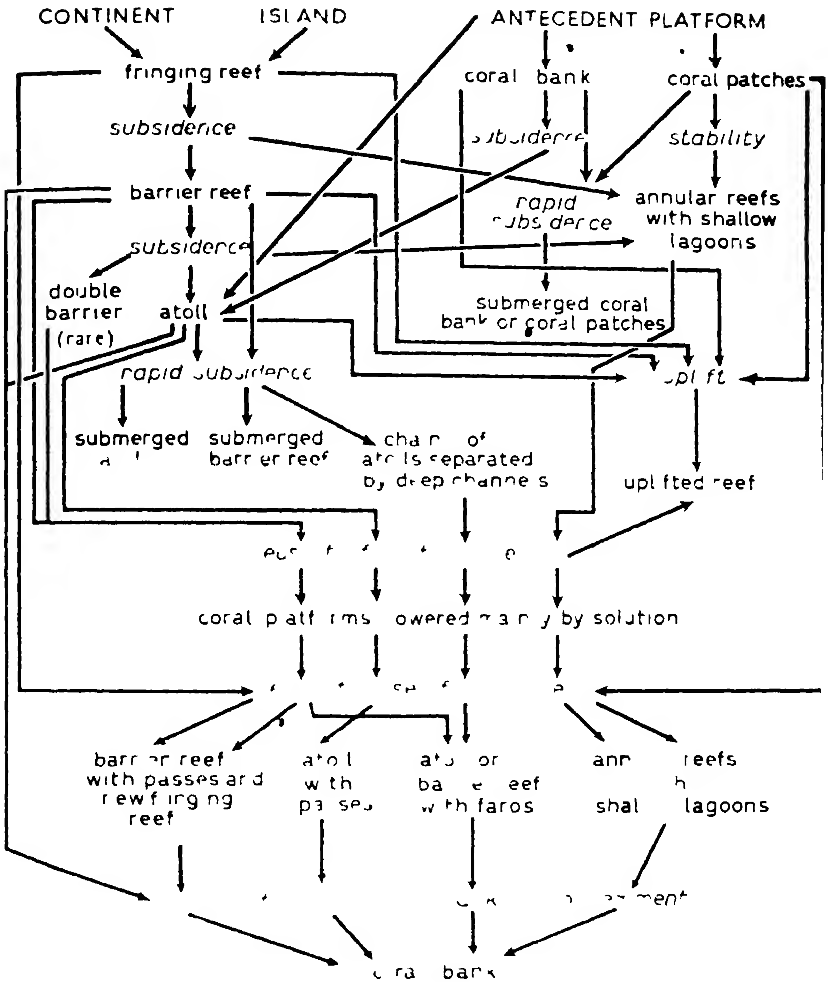
A Partial section through an ocean atoll (after various authors) B Section through fringing reef at Ras G·h··, near Jidda Red Sea (after Guilcher) Solution is affecting the dead raised reef C. Diagram of the growth curve of a coral reef (modified from Davis) D_1–D_6 Possible evolution of reefs (modified from Kuenen, 1947, and Stearns, 1946) in marginal areas where coral could not live during glacial periods with low sea-levels D_1–D_2 Tertiary D_3–D_6 Quaternary Basement rocks in white, reefs in black, talus dotted, lagoon deposits horizontal dashes

1 Submarine mount with fringing reef developed after slow uplift 2 Subsidence of mount, after partial erosion, with development of atoll, rising mount to right 3 Glacial fall in sea-level and associated coastal and subaerial erosion Influence of wind direction 4 Inter-glacial rise in sea-level and associated reef building 5 Glacial fall in sea-level and rapid subsidence of mount to right—erosion of island to left but not of one to right 6 Post-glacial rise in sea-level New atoll, enclosing a relic of the right side of the atoll shown in 4 Mount to right too low for coral growth, but slow increase of deposition may lead to conditions favourable to coral growth at later date The Tertiary stability of sea-level is hypothetical.

subsidence has been dominant in the central or 'true' Pacific, and uplift in the western Pacific west of the Andesite line (p. 239). Earth movements were of major importance in the Tertiary and possibly continued in the Quaternary, when, however, eustatic movements assumed a dominant role. At times of low sea-level, earlier coral structures were undoubtedly attacked, chiefly by subaerial solution and marine corrosion at that sea-level. Kuenen states that the channels through barrier reefs and atolls very rarely have a depth greater than 100 m. and he regards this as an effect of eustatism. In the Tertiary, the growth of barrier reefs and atolls would have been continuous, subsidence being generally slow enough for reef building to keep pace with it. The last Quaternary eustatic transgression was more abrupt, and the corals were locally killed off when the sea-level rose faster than they could build up. This may account for the channels in the reefs. Eustatism accounts for the common association of barrier and fringing reefs. It can also explain the faros as post-glacial features developing as annular forms from the Australian type and not linear forms on earlier barrier reefs, because these barrier reefs were wide enough to be subdivided, and to enclose lagoons wide enough for the water on the landward side to be in movement and well oxygenated. But the larger atolls of the Maldive complexes cannot be so explained, because the channels between such atolls and faros are too deep. A rapid pre-glacial subsidence in the area may have had the same effect as the later post-glacial transgression which gave rise to the faros (Fig. 18 D1–D6).

Conclusion. We have by no means clarified all the problems presented by reefs. A tentative diagram has been drawn (p. 135) to show some of the sequences and probable combinations but not all. In particular, the combined effects of eustatic movements and earth movements have been omitted. It should not be assumed from the diagram that in all cases evolution necessarily ends in a coral bank, as certain forms probably reach equilibrium at one of the earlier stages.

Among the unsolved problems we must mention that of the difference between the true atolls and the annular reefs with a shallow lagoon. Generally the latter are found nearer the land and the former in the open ocean. But it is difficult to account for this distribution, while in the Red Sea and in Indonesia reefs of both types occur. In both cases the exact effects of swell and currents remain to be determined. These factors are undoubtedly important, but how do they fit in with eustatic oscillations and subsidence? Can they cause the reef to migrate on its foundation for any distance, or do they effect a balance between growth


CONTINENT
ISLAND
ANTECEDENT PLATFORM
fringing reef
coral bank
coral patches
subsidence
stability
barrier reef
rapid
annular reefs
with shallow
lagoons
double
barrier
(rare)
atoll
submerged coral
rapid subsidence
submerged
submerged
barrier reef
separated
uplifted reef
coral platforms
lowered
by solution
barrier reef
with passes and
new fringing
reef
with faros
shal
lagoons
reefs
bank

and migration, as seems to be the case on the outer windward slope of Bikini atoll? Can atolls, or some of them, be an inheritance from subaerial erosion forms, controlled by limestone solution during periods of low sea-level, as Japanese scientists and Stearns MacNeil recently suggested? There are also the problems of the relative importance of corrosion and mechanical abrasion in the lowering of old reefs, and of the exact importance of this lowering. We must hope that the interest shown by the petroleum companies in reefs will result in more numerous borings, which will be complementary to submarine exploration. Such research will doubtless create problems of which we are as yet unaware. Coral reefs form one of the most complex questions in geomorphology.

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Chapter IV

CLASSIFICATION OF COASTS

Having dealt with the forms produced by the sea, we may now proceed to the problem of the classification of coasts, in which we shall have to take into account also the 'initial' (Gulliver) or 'primary' forms (Shepard),[1] which are, of course, not of marine origin.

Coasts, on which the majority of forms are of the type described in the previous chapter, have obviously been strongly subjected to the influence of the sea. The length of time required for a coast to obtain such a strong marine imprint varies with the nature of the rocks, the height of the coast, and the degree of activity of the marine processes. Each 'initial' type evolves into a 'sequential' form. The initial types will be therefore described and brief indications given of the modifications which the sea causes in each case. An examination of coastal evolution in general is given in Chapter V.

A. RIA COASTS

Rias may be defined as river systems partly or wholly flooded by the sea. The degree of drowning depends on the magnitude of the movement of base-level and on the altitude of the source of the river. The subaerial origin of rias is demonstrated by the occasional existence of incised meanders as on the Aulne at Landevennec in the Rade de Brest. In our opinion, it is not necessary, however, to restrict the term rias to deeply incised drowned valleys: we may, for example, use the term ria for Chesapeake Bay, and the drowned valleys between Pont-l'Abbé and Penmarc'h in south-west Brittany. Ria coasts include then all those possessing a great number of flooded river valleys. The relation of the total length of the coast to the length of a smooth line through its headlands is, therefore, very high. Usually flooding is the result of the Flandrian transgression, but may also result from subsidence, especially in the

[1] *Initial* seems to be much better than *primary*.

Pacific islands, as for example in New Caledonia according to Davis. The characteristics which distinguish ria from fjord coasts will be examined later.

Ria coasts are widespread. Those of Brittany, where the ria is often called an *aber*, are well known from the articles of de Martonne (Fig. 19 A). Those of Galicia and Shantung are equally well known. But similar coasts also occur in south Cornwall, south Wales, south Ireland, south China, Korea (Fig. 19 C), between New York and Washington, at the mouths of the Gambia, Casamance and adjacent rivers of Portuguese Guinea, between Cameroon Bay and the mouth of the Gabon, on many Pacific islands such as north-east New Guinea, New Caledonia, Moorea, Huahiné, Raiatea, and Tahaa in the Society Islands, and in the vicinity of Sydney.

The bays and headlands of ria coasts evolve somewhat differently. The bays are in reality estuaries and subject to typically estuarine deposition of mud and consequent evolution of marsh and tidal channels, and, in tropical regions, mangrove swamps. Deltas are sometimes formed when much sediment is brought in by river. Examples of very advanced silting are provided by the Gambia, Casamance and adjacent rivers. In Tahiti silting is almost complete. This evolution involves the decrease in size of the channel, which has been caused by submergence and hence is too wide for the inflowing streams.

The headlands tend to be cliffed, but as the Flandrian transgression is so recent the cliffs are often no more than small notches at the base of the old valley sides. These notches usually, but not always, occur in exposed areas: some occur inside the bays, where the latter are at least 150 to 200 m. wide at high water. In these bays therefore there is erosion at high water level and muddy sedimentation lower down. This is the case in Brittany, where the periglacial deposits forming the slopes are cliffed in all the rias.

The principal change affecting drowned coasts is the formation of sand or shingle spits of various types, either at the mouths of the rias or between their mouths and their heads. Bay-mouth bars on the east coast of the United States are well known, e.g. Cape Hatteras; in Brittany, the V-shaped spit of Mousterlin south of Quimper dams the small rias of the Fouesnant region (Fig. 7 A and B). Although such forms smooth out the general plan of the coast, they also increase its length as long as the lagoons behind the spit are not infilled. As the lagoons are filled up a marsh is formed, as in the south-east of the United States.

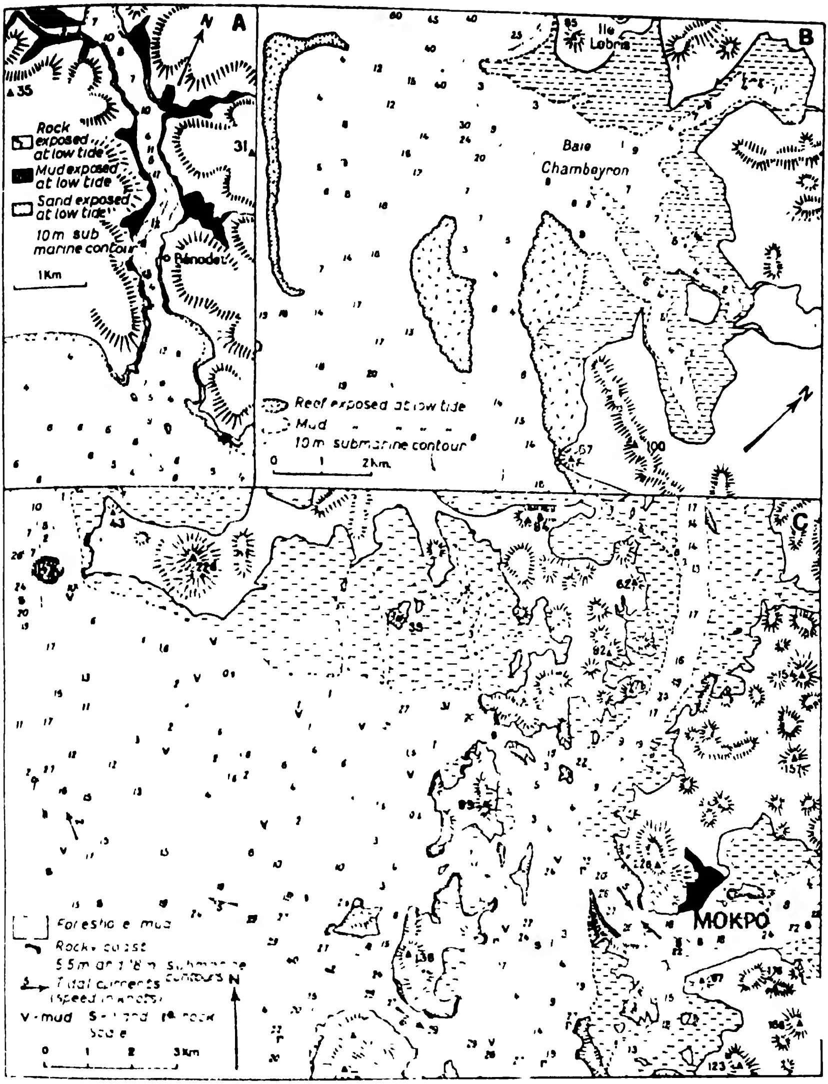

FIG. 19 RIA COASTS

A. Lower part of Odet ria, Brittany, based on French marine chart 5368. Note scour holes, deposition at mouth of estuary and distribution of sediments at the sides of the ria. B. West coast of New Caledonia in 21° 50′ S., based on French marine chart 3806. Estuaries partly cut off by discontinuous barrier reef and silted up. Mud is spreading on to the fringing coral reef. C. South-west coast of Korea in 34° 48′ N. after Admiralty chart 3392, which is based on Japanese charts. Note typical drowned coast, extent of mud-flats, violence of tidal currents in straits, and numerous scour holes. All heights and depths in metres.

In coral regions, the function of these spits is taken over by the barrier reefs which have grown up during submergence. The Pacific islands mentioned above usually have barrier reefs, especially New Caledonia (Fig. 19 B). The inner shore often has a fringing reef, which may extend some way into the rias if the streams flowing into them bring little alluvium, so that silting is, for some reason, either unimportant or absent.

A special type of ria is the *sherm*, found on both coasts of the Red Sea (Fig. 20 A). They are long bays with a narrow entrance, cut in the low

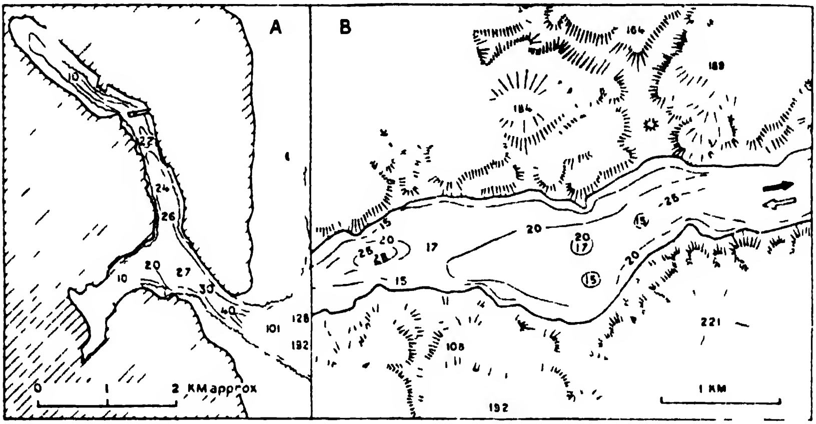

FIG. 20 SHERM AND DROWNED TUNNEL-VALLEY

A. Port Sudan sherm (after Rathjens and von Wissmann, 1933). Depths in metres. B. Drowned tunnel-valley at Mariager, in North Jutland, in 56° 40′ N., 9° 50′ E. (after Schou, Atlas of Denmark, 1949). The black arrow points seawards and the white arrow gives the direction of flow of meltwater in the Ice Age. The lateral ravines, which are not drowned, are post-glacial. All heights and depths in feet.

coastal plains, and often widening or branching within. Some, for example the Ubhur sherm 27 km. north of Jeddah, have the appearance of drowned meanders; the depth, which is usually several dozen metres at the entrance, decreases gradually towards the head. There is not always a wadi at the head of the main arm, and often none at the heads of the branches. They are excellent sites for harbours; e.g. Suakin and Port Sudan.

It is not certain that sherms are rias in the genetic sense of the term. The hypothesis is tempting when they meander and continue the line of a wadi. But as they are bordered by slightly raised coral reefs and as

coral still lives in those where harbour works have not prevented it, Rathjens and von Wissmann have suggested that they were bays enclosed by the growth of coral, the general direction and branching of the sherms being caused by structural control. This interpretation, which was suggested by the analogy with the bottle-shaped bay of Farsan Kebir Island in the Red Sea, is put forward with some reserve. Perhaps there are sherms of both tectonic and ria type. Even in the latter tectonic effects may be important, as suggested by Schmidt.

B. FJORD COASTS

Fjord coasts like ria coasts are fundamentally due to submergence, since fjords are glacial valleys flooded by the sea. However, fjord coasts, like many glacial coasts, have undergone a post-glacial isostatic uplift, which in certain places such as the west coast of Norway was certainly greater than the Flandrian transgression, so that such coasts are, in one sense, emergent coasts. They are, however, predominantly submergent in form. because the glaciers which occupied the present fjords were so thick that they could not float in the sea until they reached deep water. In Norway, if we assume that the extent of the Flandian transgression was 100 m., and the isostatic movement of the land at least 200 m., the glaciers had every chance of deepening their valleys hundreds of metres below the present uplifted pre-Flandrian level. These facts satisfactorily account for the enormous depths found in certain fjords, the greatest being in the Sognefjord (1,244 m.), in the Messier channel in Patagonia (1,292 m.), and Scoresby Sound in Greenland (about 1,300 m.). Fjord coasts were formed by valley glaciers cutting down well below sea-level and drowned, not by the Flandrian transgression, but by the disappearance or shrinking of the glaciers.

We must define fjords, because all valleys in glaciated regions, even hilly ones, are not fjords, neither are submerged closed depressions proof that the valley is a fjord.

There are true rias, that is drowned river valleys, in formerly glaciated areas: for example in Wales, and the south of Ireland. The mouths of the Towy in Carmarthenshire, the Conway, north of Snowdon, and the Cork river are rias. In these regions there are no typical glacial erosion forms, either because the glaciers were insufficiently powerful or, more generally, because the pre-glacial valleys were not sufficiently steepened by rejuvenation to lend themselves to the formation of glacial troughs. A youthful or rejuvenated relief is, therefore, essential. Johnson has

Scandanavian authors, is that of the *Strandflat.* This is (Fig. 21 B) a coastal platform, varying from a few kilometres to 60 kilometres in width, on the Norwegian coast, especially in the west and north-west and ending abruptly inland against the mountains. It is partly below and partly above water and includes the innumerable reefs of the *skjaergaard.* Nansen distinguished three levels: + 30 to + 40 m., + 15 to + 18 m.; and about 10 m. It is crossed by channels, which are the continuations of the fjords. The depth of these may reach several hundred metres but is never as great as in the great inner deeps of the Sognefjord. The cross-section of these channels is the same as that of glacial valleys. Strandflats also occur in Bear Island, Spitzbergen, Iceland, and west and south Greenland (Nansen, O. Holtedahl, Werenskiold, Hjulström). The strandflat appears therefore to be a form frequently associated with fjord coasts and not one peculiar to Norway. Ahlmann (1919) considers the Norwegian standflat to be an old subaerial peneplain of earlier date than the trough-like channels which cross it. Nansen, on the other hand, thought it to be a coastal surface, partly initiated by the ice-foot (pp. 25–6) and later developed by the action of waves in its exposed parts. It is thought to be later than the troughs which cross it, since it continues as a shelf or bench inside the fjords, notably in the Sognefjord and the Hardangerfjord. It would have been modified and possibly extended during the interglacial periods, especially during the cold periods immediately preceding each advance of the ice. The three levels of the strandflat may thus correspond to various periods of modification and, according to Nansen, suggest that there were three interglacial periods in Norway. In the glacial periods the strandflat would have suffered some erosion and partial destruction.

More recently (1946) E. Dahl has proposed a different explanation. The existence of a relict alpine flora in Norway suggests that the Fennoscandian ice-sheet did not cover all the coastal regions of Norway. The summits of the Lofoten Islands were above it, because the thickness of the ice there was reduced by the rapid flow caused by the breaking off of icebergs along the neighbouring continental slope. Corries developed, and ultimately produced a smooth surface as their recession destroyed the highlands: this surface is the strandflat. Farther south, where the continental shelf is wider, the strandflat was produced by the ice-sheet smoothing out areas, already partly dissected by corrie glaciation. But this explanation did not receive universal acceptance in Norway, and in 1955 H. Holtedahl produced morphological evidence against Dahl's botanical arguments.

The fact that strandflats are characteristic only of fjord coasts, seems to mean that their formation ought to be attributed in large part to glacial or periglacial processes, but perhaps also to others, as Hjulström has suggested for south-east Iceland.

Since the ice retreated, fjord coasts have been only slightly modified by marine or river action. Small alluvial plains have been formed at the heads of the fjords, e.g. in the west of Scotland, if we may assume that these are true fjords. Ice erosion continues at sea-level, where freezing is still common. But the filling up of fjords is very slow, for they are much deeper than rias.

C. GLACIAL LOWLAND COASTS

Glacial plain coasts may be emergent coasts, as in Sweden, where the post-glacial isostatic rise still continues; but they may also be coasts of submergence where the relief, due to glaciation, has been drowned by the Flandrian transgression and the melting of the ice-sheet, which has been replaced by the sea. In Sweden we may speak of submerged glacial topography now in process of emergence.

These coasts are as intricate and characterized by as many islands as the fjord coasts. They are morphologically much more varied because they have in juxtaposition so many glacial accumulation forms: terminal moraines, drumlins, outwash plains, proglacial channels, subglacial valleys, eskers, kames, and kettleholes. Finally, the glaciation of a low-lying peneplain may produce subdued structural landforms. The relative dominance of each of these forms gives a distinct appearance to the coast.

In New England Johnson has noted characteristic types. Two terminal moraines parallel to the general trend of the coast in the Nantucket and Cape Cod region have lagoons behind them. The drumlins of Boston harbour tend to give rise to oval islands and peninsulas (Fig. 26 A). Outwash plains are marked by broad lobes cut by old slightly-incised, proglacial channels, which have been drowned by the sea, as at Martha's Vineyard (Fig. 26 B). Eskers appear as sharp points, as at Weymouth. Kettleholes form small enclosed bays in the neighbourhood of Cape Cod.

The same forms recur on the east coast of Hudson Bay where a slightly uneven peneplain dips gently under the sea. The coast is bordered by numerous low islands. Similar conditions occur in the north of Labrador, and in other Arctic regions of Canada. The Baltic area is extraordinary rich in this respect, e.g. in Finland, Sweden, where

the channels are called *fiärds* (Fig. 21 C), Poland, Germany, and Denmark. In these last two countries there are forms similar to those of New England as well as coasts with drowned subglacial valleys, or *föhrden*, which were orginally cut by subglacial torrents flowing towards the south, south-west, and west (Fig. 20 B). The bottoms of the *föhrden* are flat and may exceed 1 km. in width, while the valleys are sometimes incised to a depth of more than 100 m. The slopes are steep and contrast sharply with the evenness of the moraine plateaux. The essential difference between *föhrden* and fjords lies in the shallow depth of the former and the absence of closed deeps.

The isostatic rise of the land has often left a succession of old shorelines, which add another element to the landscape (Chapter II). In Denmark the low dead cliff of the Littorina sea commonly occurs just behind the present shore. Raised beaches are more numerous farther north in Scandinavia, but they are not peculiar to glacial lowlands, for they may be found on fjord coasts.

An extremely interesting example of the complexity which can occur, is provided by the Stettiner Haff, which has been thoroughly studied by Braun and Uhden (Fig. 21 D1–D3). This Haff is situated in a basin formed to the south of the recessional moraine of the Baltic ice-sheet, when the latter was in the region of Usedom and Wollin islands. The moraine still forms a part of the land enclosing the Haff. The melt waters then flowed southwards, by channels which are now abandoned or almost abandoned valleys (Uecker and Randow). The basin occupied by the Haff has been partly filled with fluvio-glacial waste, then by the sand terraces of the pro-glacial valleys, and finally by wind-blown sand. It was then invaded by the Littorina sea, and a system of spits with recurved ends constructed by the sea. These have now completed the enclosure of the Haff near Swinemünde (Swinioujscie).

The Stettin Haff leads us to the evolution of a glacial lowland coast as a result of marine action. This evolution has reached a more advanced stage than that of the fjord coasts since the deposits are unconsolidated, easily attacked, and redistributed by the sea. In both North America and the Baltic, glacial lowland coasts show rapid evolution and a remarkable development of spits of all kinds. Examples as good as those near Swinemünde are found on the Zealand coast, described by Schou, and on the American coast at Nantasket and Cape Cod, studied by Johnson and Davis. As the moraines are eroded, the fulcrum of the spit disappears and the older beach ridges are eroded by the sea. There are no better places in the world where one may see and

appreciate the trend of shoreline evolution. This evolution undoubtedly leads to a simplification and straightening of the shoreline (see Chapter V, pp. 180–2).

D. UNGLACIATED LOWLAND COASTS

Unglaciated lowlands have regular shorelines and thus contrast with glacial lowlands. They are diversified only by estuaries and deltas, in the evolution of which the sea plays a part. As a result, sequential forms (see Chapter III) predominate, so that little need be said about them here. The Flandrian transgression has affected all these lowlands. Where deposition is active on these coasts, the constructive action of the sea may lead to the building of offshore bars and other features; if a marked submergence affects them, it is usually the result of subsidence. In the Flemish and Dutch plains, deposition roughly balances subsidence and there is approximate stability.

The characteristic form of this type is the offshore bar enclosing a lagoon behind it. Examples occur in Languedoc, on the north coasts of the gulfs of Mexico and Guinea, and, on a smaller scale, in the Harwich area north of the Thames estuary. The coast of Gascony is of the same type, but has reached a later stage of evolution.

Unglaciated lowland coasts are often marshy coasts, the marsh lying either behind an offshore bar and being caused by silting of the lagoon, or at the edge of the sea, e.g. the marsh of Dol near Mont-Saint-Michel. The best coasts of this kind are along the North Sea coast from Flanders to Jutland, and in Texas, but others are found in Guiana, and in the east of Madagascar, especially near Tamatave, where they are prograded seawards by the construction of successive beach ridges.

Periglacial lowland coasts occur mainly on the Arctic coasts of Siberia and America and are remarkable for their deltas which are drastically changed by each annual flood (p. 113). They only develop in summer, because the sea and rivers are frozen during the rest of the year. On the north coast of Alaska blocks or masses of tundra, bounded by vertical ice-wedges, break off partly as a result of marine action, and partly through the melting of the ice.

E. COASTS DOMINATED BY STRUCTURE

In the ria and fjord coasts, structural influences often greatly affect the outline of the coast. It may not be easy to decide whether the characteristics are due to submergence or to structure: the Rade de Brest is a

typical ria coast, but the rias follow beds of soft rocks. In other cases structural influences are so important that they are the only possible basis for classification. We shall separate coasts with longitudinal, transverse, oblique, arcuate, rectangular, and discordant structures. In some of these types direct and indirect structural effects and lithological effects may be distinguished. Finally, we shall consider volcano coasts.

Coasts with longitudinal structure

These coasts may be called Pacific or Dalmatian, but they are not absent from the coast of the Atlantic.

Direct tectonic influences. If one agrees with Bourcart and Jessen (Chapter II), the great majority of coasts would have to be considered of this type, since their outline would be determined almost everywhere by tectonic movements parallel to the coast. Even if one does not adopt these views, it must be admitted that direct tectonic effects often produce coasts where the component parts of the coast are parallel to the coastline as a whole.

The simplest type is the coast formed by a recent fault, especially if the fault scarp is rectilinear. Remarkable examples are found in the Gulf of California, particularly on the south-west of the island of Angel de la Guarda (Fig. 22 A), where the escarpment is continued by steep submarine slopes leading to a trench with a flat bottom at a depth somewhat in excess of 1,000 m. In the same region, another example is provided by the west coast of the island of Santa Cruz (Shepard). Cotton has described examples of this type in New Zealand, notably at Port Nicholson near Cook Strait. The coast of Queensland also appears to be of the same type in many places: the Great Barrier Reefs may be built up on a foundered area at the foot of a fault and limited on the east by a similar fault scarp entirely below sea-level. When the fault is very recent plunging cliffs may occur. According to Cotton, these resist erosion because they reflect the waves since there is no shallow platform at their foot to cause the waves to break.[1] But the presence of plunging cliffs is not in itself proof of a fault-scarp coast: they may result, for example, from the Flandrian transgression on a coast of high cliffs of quite different origin. The sides of fjords are false plunging cliffs of another kind.

[1] The total absence of a platform would therefore have the same effect as a very broad platform, which weakens the waves before breaking and consequently minimizes their mechanical effect.

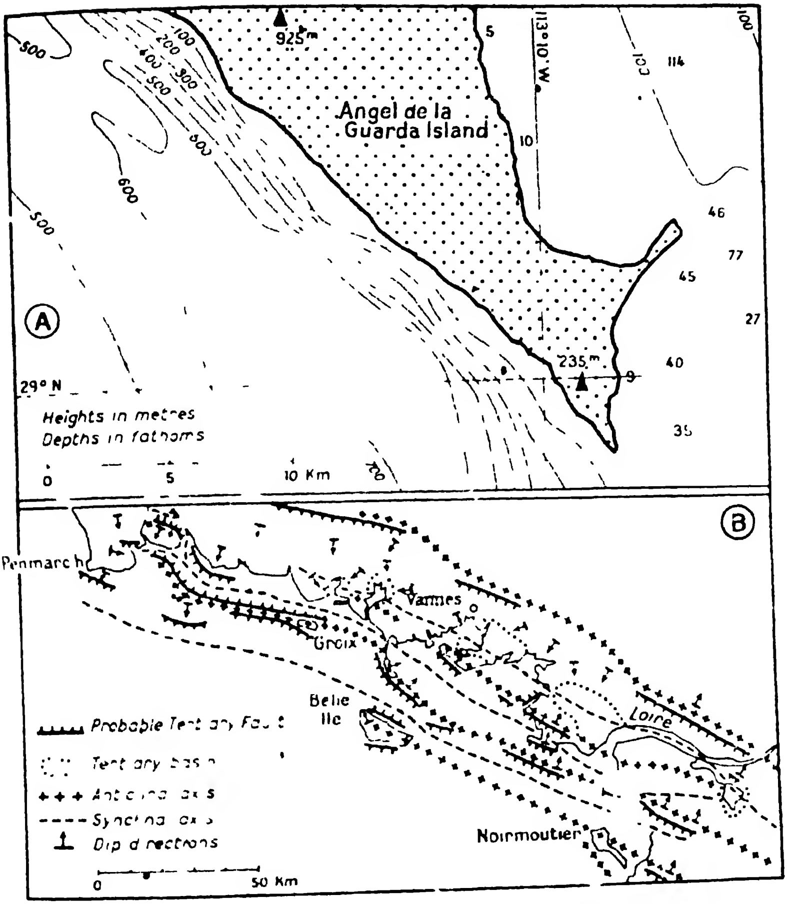

FIG. 22 COASTS WITH LONGITUDINAL STRUCTURES

A. Faulted coast with plunging cliffs on the south-west side of Angel de la Guarda Island in the Gulf of California, Mexico (after Shepard, 1950). Compare the depths off the two sides of the island. B. Tectonic effects on the coast of south Brittany (slightly simplified from Guilcher, 1948).

A recent tectonic movement may be inferred when a shore presents certain other abnormal characteristics. According to Cotton, it is curious that the Makara river in the Wellington peninsula in New Zealand does not enter the sea in a ria; the anomaly can be explained by a recent uplift of the coast near its mouth.

When a series of faults runs parallel to the coast, upfaulted blocks may remain as islands in front of the coast. According to Chardonnet, the Provençal coast between Cape Sicié and Cavalaire is of this type: fault scarps or fault-line scarps are presumed to border a large part of the continent: other fault scarps appear to form the southern limit of the Giens peninsula (formerly an island) and the islands of Porquerolles and Port-Cros. A coast may be influenced by both folding and faulting. Such appears to be the case in southern Brittany between Penmarc'h and the Vendée (Fig. 22 B): the islands and archipelagos correspond to Tertiary anticlines, faulted in several places. The same structure continues inland, where the downfaulted parts have been submerged to form an aligned series of inlets; Pouldon Bay (Pont l'Abbé), Kerogan Bay (Quimper), Rade de Lorient, Etel river, East Morbihan, Traict du Croisic (Fig. 7 C), Brière.

Southern Brittany reminds us of the classic example of Dalmatia, where the longitudinal arrangements of islands and promontories correspond to anticlines and the channels and bays, such as that of Kotor, to synclines. It is immaterial whether this is simple submerged relief or whether, as Bourcart believes, the anticlines and synclines are still in course of formation. At any rate the coast is very young and there are hardly any cliffs, as the sea has had insufficient time to form them.

Indirect tectonic and lithological influences. Ancient faults may also guide the work of the sea. In southern Brittany, between Concarneau and Le Pouldu, crush-zones roughly parallel to the shore have enabled the sea to cut the coast into small rectilinear masses (Fig. 3 C).

In an old massif of Appalachian-type folds, if the rock outcrops are parallel to the coast, and former river valleys are drowned, the relief is like that of Dalmatia but of different origin. Good examples of this occur near Cork, where the wider parts of the harbour coincide with beds of Carboniferous Limestone and the narrower parts with Old Red Sandstone (Fig. 23 A); in north Brittany in the Lézardrieux river which is broader in the Plounez schists than in the Plourivo sandstone upstream and the Paimpol spilite downstream; and in the Rance estuary at Saint-Suliac and La Ville-ès-Nonais (Fig. 23 B). In North America, Johnson has quoted very good examples of the same type in the east of

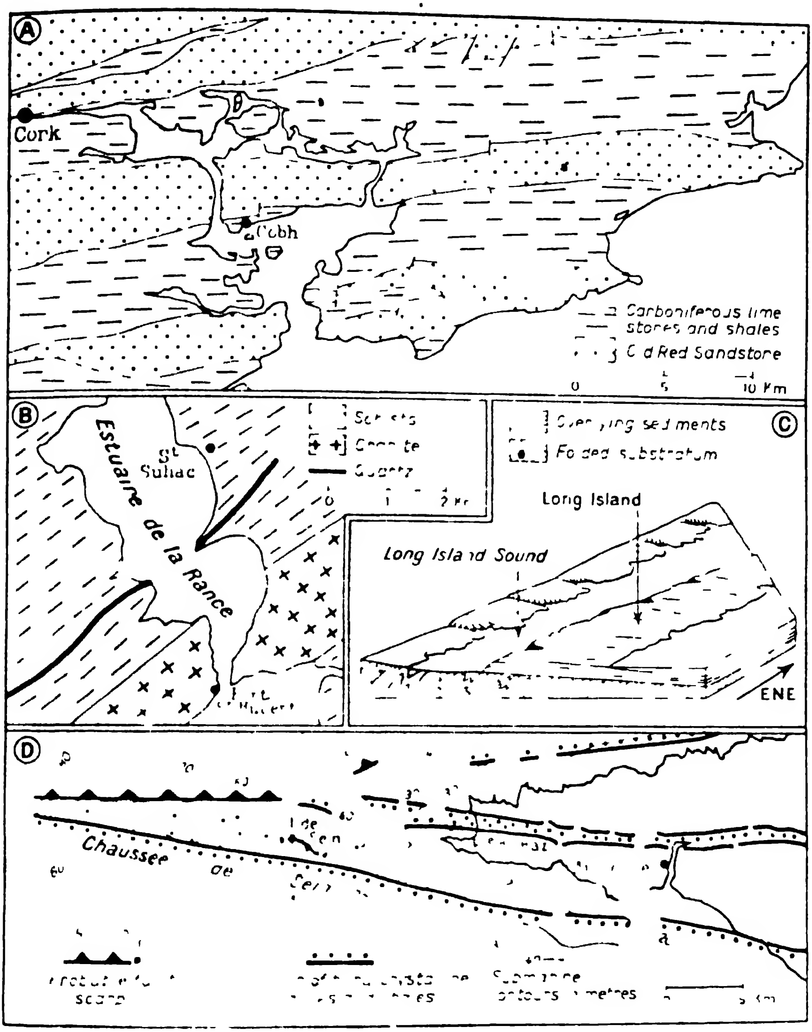

FIG. 23 COASTS WITH LONGITUDINAL AND TRANSVERSE STRUCTURES

A. Cork Harbour, Ireland. B. Rance estuary, north-east Brittany. C. Long Island cuesta, New York, and submerged depression (after Johnson, 1925). D. Pointe du Raz and Chaussee de Sein, south-west Brittany (after Guilcher, 1936 and 1948).

Cape Breton Island, where drowned depressions occur in soft Carboniferous rocks, but are narrowed by crystalline rocks, and along the St John river before its outlet into the Bay of Fundy; further to the south he mentions also the drowned peripheral depression between the escarpment of Long Island and the mainland (Fig. 23 C). Thus the submergence of scarplands can give the same type of form as the submergence of Appalachian relief.

Transverse coasts

These coasts may also be called Atlantic coasts. The continental structures meet the coast at or very nearly at right angles.

Direct tectonic effects. Recent folds or faults sometimes affect the general trend of the coast. Where the Great Atlas Mountains meet the Moroccan coast the coast projects, while to the south, the Sus syncline, which has suffered movement in Late Pliocene times, forms a re-entrant. Within the Atlas anticlinorium, the anticlines of Tafelneh and Cape Rhir form headlands, which project more sharply on their southern sides because the anticlines are asymmetrical with steep dips on their southern limbs. In Morocco again, abrupt re-entrants south of Cape Cantin and Cape Blanc near Mazagan are perhaps due to faults perpendicular to the coast, the throw to the south-west or south having carried that side below sea-level. The promontory of the Chaussée de Sein (Finistère) with its steep northern slope seems to fall in the same category (Fig. 23 D). Another example is mentioned by Cotton near Napier on the east coast of the North Island of New Zealand: recent folds perpendicular to the shore, some of which are still in process of movement, give rise to cliffs on the anticlines and low coasts on the synclines.

Indirect tectonic and lithological effects. The examples are comparable with those of longitudinal coasts. In the north-east of Nova Scotia (Johnson, *New England*, p. 41), valleys were cut along major faults perpendicular to the coast, and are now submerged giving rise to long estuaries. On the north-west coast of Scotland, Loch Ewe and Loch Maree similarly follow an important north-west–south-east displacement (Fig. 24 B).

Ireland and west Brittany afford good examples of drowned transverse structures of Appalachian type: on the south-west coast of Ireland the bays are in the Carboniferous Limestone and the headlands in the Old Red Sandstone; on the east coast of the Rade de Brest the bays are in Middle Devonian shales and the headlands in Lower Devonian

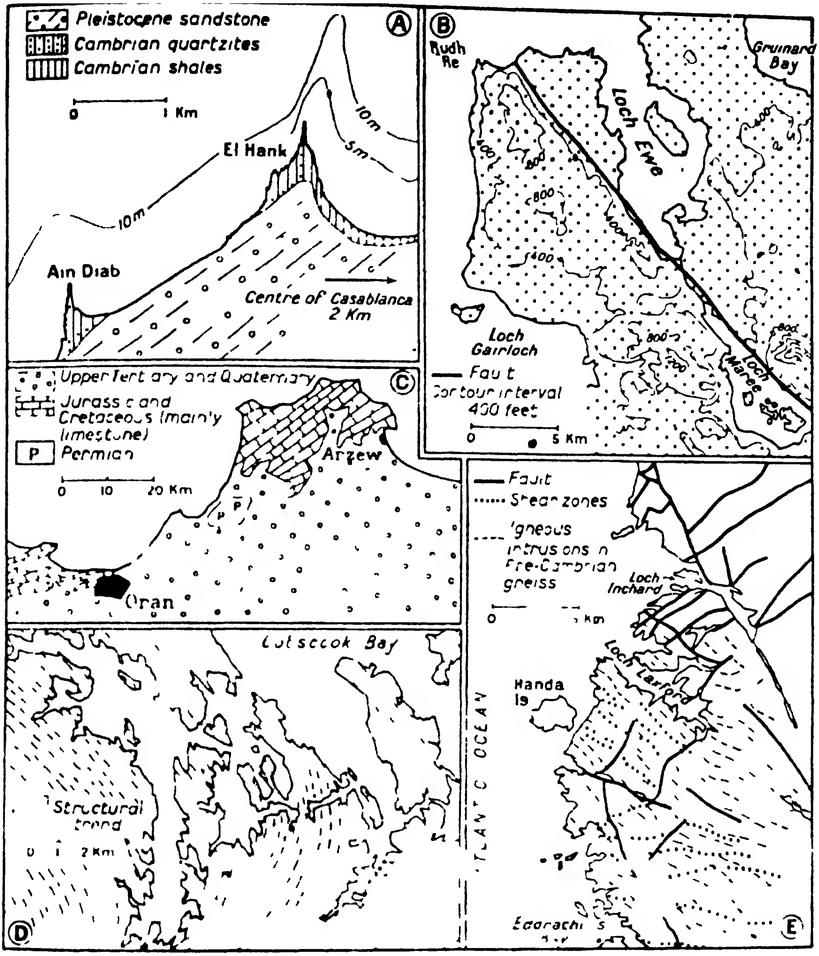

FIG. 24 COASTS WITH TRANSVERSE, OBLIQUE, ARCUATE, OR RECTANGULAR STRUCTURES

A. Transverse structure at Casablanca, Morocco. B. Transverse structure on west Scottish coast near Loch Ewe, Wester Ross. C. Oblique structure at Cobscook Bay, near Eastport, Maine (after Johnson, 1925). The rocks are interbedded shales, sandstones, and volcanics. E. Rectangular structure in north-west Scotland: the north of the area shown is 14 km. south of Cape Wrath.

flagstones and quartzites. In Morocco, the *skhour*, which are small quartzitic crests orientated north-south on the surface of the Meseta, form small headlands on the coast near Casablanca, e.g. El Hank to the west of the town (Fig. 24 A).

Coasts of oblique structure

Direct tectonic effects. In Algeria the folds of the coastal ranges run north-east-south-west, while the trend of the coast is east-north-east-west-south-west. The result is a series of oblique headlands corresponding to the anticlines or anticlinoria with bays, e.g. Oran, Arzeu, and Algiers, in the synclines, which are well protected from west winds by the neighbouring anticline, but completely open to the north-east (Fig. 24 C). This is clearly an example of direct tectonic effects, for many of the folds of the Algerian coast have undergone movement in the Quaternary and even in modern times. But lithology plays a large part in preventing rapid erosion of the anticlines, since they are formed of Mesozoic limestones or of old rocks, which are much more resistant than the beds in the synclines.

In Morocco, old consolidated dunes are also cut obliquely by the shore at Fedala; oblique headlands correspond to lines of dunes, and re-entrants, e.g. the harbour at Fedala, to the troughs. This is not a tectonic effect, but the result is the same.

Indirect tectonic and lithological effects. Near the Bay of Casco in the State of Maine (Johnson, *New England*, p. 13), partly submerged Appalachian relief runs north-east-south-west oblique to the east-north-east-west-south-west coast. The landforms produced by the great Glen More fault in Scotland and the Norwegian coast near Trondheim may be included in the same class.

Coasts of arcuate structure

Occasionally concentric arcuate folds form a series of concentric bays connected by straits, as in the flooded Appalachian structures of the Cobscook Bay region of Maine (Fig. 24 D) and in similar structures in Norway near Bergen.

Coasts of rectangular structure

These coasts are common, especially in old rejuvenated massifs: the effects are, therefore, indirect. It has been seen earlier in this chapter that fjord coasts often show networks of intersecting faults which have guided river and later glacial erosion. Even though the north-west coast

of Scotland may not be a true fjord coast, it is greatly affected by intersecting structural trends, notably between Rhu Coigach and Loch Inchard, where alternating north-west–south-east and north-east–south-west sections correspond respectively to zones of intrusion or shearing in the Lewisian gneiss and the Torridonian conglomerates, and to faults affecting the same rocks (Fig. 24 E). The north–south trend, seen to the south-east of Rhu Coigach, is due to faults affecting the Cambrian as well. The coast of the Orkneys also provides examples of rectangular structure: one such series of faults has given rise to the harbour of Scapa Flow. The south coast of Norway, near Kristiansand, presents an example of the effect of erosion on a very fine network of faults in the pre-Cambrian massif. The Gulf of Morbihan, which as a whole is part of the longitudinal coast of southern Brittany, shows in detail some rectangular features caused by small granulitic massifs perpendicular to the general strike, e.g. the Ile aux Moines. The coast of the Inland Sea of Japan falls in the same class since the general trend, according to F. Ruellan, depends on dislocations parallel to the main longitudinal folds, but broad transverse upwarps give rise to headlands on the anticlines and bays on the synclines. Finally, the west coast of Asia Minor is a very good example of the direct effect of a rectangular structural pattern, especially the Khios and Izmir area.

Coasts of discordant structure, or contraposed coasts

Strictly speaking, the Long Island example already mentioned must be classed with coasts characterized by discordant structure, since it includes a peripheral depression But most of the coasts in this category have unconsolidated deposits such as glacial or periglacial drift, including loess resting on much more resistant rocks. Su arrangements lend themselves to the formation of contraposed coasts.

Thus defined, contraposed coasts chiefly occur in formerly glaciated regions and in places where the periglacial deposits are thick, e.g. the loess on most of the north coast of Brittany from Dinard to Ploudalmézeau. General and detailed features must often be considered separately in classifying them: the coast of Trégorrois and Léon is a ria coast in general outline, but a contraposed coast in detail. On contraposed coasts the drift usually covers, at least in part, a very rugged topography in the hard rocks below; thus, the drift varies greatly in thickness. In north-west Wales near Pwllheli and Aberdaron (pl. VIII B) high hills of igneous rock emerge from a mantle of drift. These hills, together with pre-Cambrian gneiss and Ordovician slate, form the skeleton of the

coast while the drift occurs only in the bays. The same thing happens in the west of the Gower peninsula where the skeleton is formed of Devonian sandstone and Carboniferous Limestone. In Trégorrois and Léon, there is a confused relief of small granite masses covered by loess, solifluxion deposits, or weathered rock (Fig. 25 A). Such coasts also occur in New England, where they have been studied by Clapp.

The evolution of this type poses the problem of differential marine erosion, which will be considered in Chapter V. The evolution of other coasts showing structural effects will also be dealt with in the same chapter.

Volcano coasts

According to Shepard, volcano coasts fall into two types.

Circular or lobate coasts. A volcanic island should have a more or less circular or elliptic outline. Many volcanic islands show such forms, e.g. Amsterdam and Tristan da Cunha in the southern Atlantic Ocean. There are also volcanic peninsulas with similar rounded outlines. Islands consisting of more than one volcano may have lobate coasts, e.g. Hawaii where Mauna Loa, Mauna Kea, Kilauea, Kohala, and Hualalai all form lobes (Fig. 25 B). The most perfect forms are produced by basalt volcanoes as the fluid lava flows regularly in all directions, but other types of volcanoes also produce circular or lobed coasts, for example, Mont Pelée, which is responsible for the curved shape of the north-west coast of Martinique.

The coast of nearly circular granite massifs may be included here, although the emplacement of the magma is affected in quite a different way. If the granite is surrounded by softer rocks the coast is similar to that of a volcano: e.g. the Land's End peninsula in Cornwall.

Caldera coasts. The origin of the calderas is outside the scope of the present discussion. It is sufficient to differentiate calderas due to (*a*) subsidence, (*b*) explosion, and (*c*) erosion. The distinction between the first two types is sometimes difficult to draw. Flooding of calderas by the sea creates a unique type of coast, with very steep slopes rising from a central mass of water: where the outer gentler slopes have not been destroyed the outer coast may be lobate.

A very fine caldera coast caused both by subsidence and explosion is that of Santorin in the Aegean, which is characterized by extremely steep slopes on the inner side and much gentler ones on the outer side (Fig. 25 C). The small volcanic islands of Palaea- and Nea-Kameni which have been formed in the flooded caldera only affect the picture

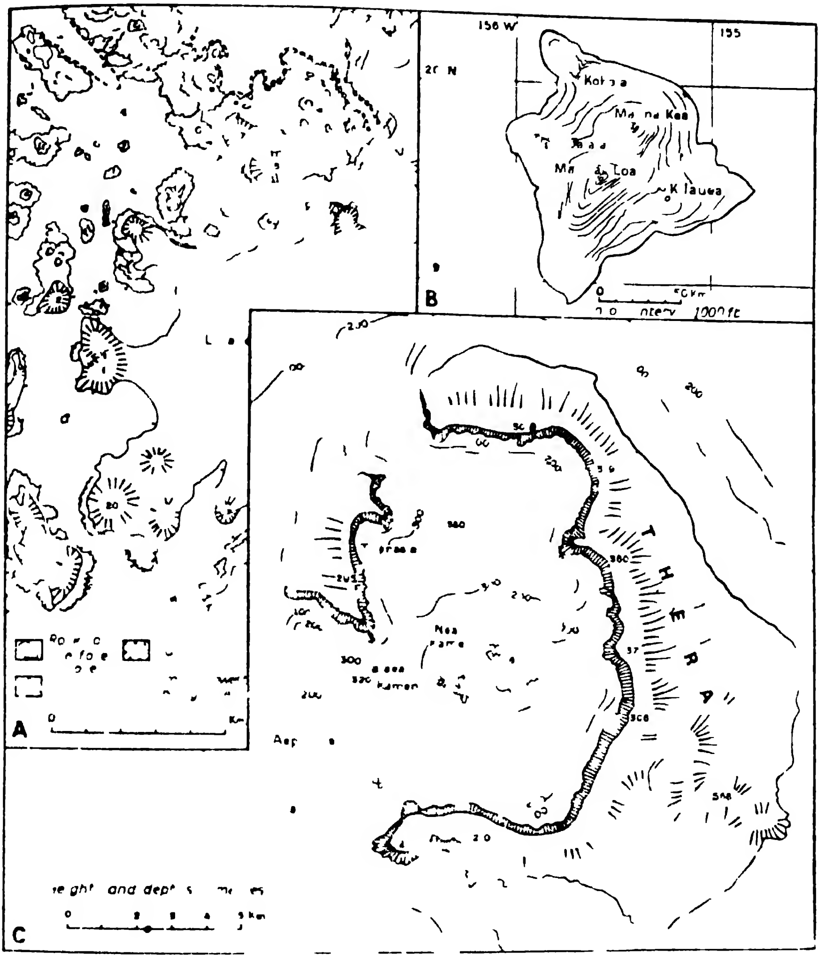

FIG 25 CONTRAPOSED AND VOLCANIC ISLAND COASTS

A Coast at I il a in Plouguerneau, Leon north-west Brittany Exhumation of granite hummocks forming rounded headlands between coves excavated in loess and periglacial deposits and, later, islands B Hawaii An island of basaltic domes C Santorin in the Aegean a caldera Note the asymmetric form with gentler slopes on the outside (the relief of the island is shown diagrammatically)

in detail. St Paul's Island in the Indian Ocean and Krakatoa afford similar examples. The island of Oahu in the Hawaiian group is formed of two half-calderas back to back and facing the sea. An example of a flooded erosion caldera is provided by Lyttelton Harbour in the Banks peninsula, New Zealand: a broad inner bay is connected with the sea by a fairly narrow channel. The bay has been interpreted as an erosion caldera by Speight, Davis, and Cotton, and the channel as a large barranco.

Originally, circular or lobate volcanic coasts had no cliffs, the slope of the cone continuing under the sea, but marine erosion may soon cut cliffs in them, thus breaking the continuity of the slope. At Tristan da Cunha the cliffs reach a height of 300–500 m. (lava cliffs are very often vertical). If rivers cut valleys in such cliffs and are later submerged a ria coast is produced. Marine erosion of the inner walls of calderas is retarded by the fact that the waves have a very small fetch unless the caldera is widely breached as at Krakatoa. Erosion is extremely variable, being very rapid in ashes and tuffs (p. 70), and much slower in rocks such as basalts and trachytes. Finally, in coral seas where volcanoes are numerous, the growth of coral introduces many complications.

Coast which do not fall into these classes may be included in those chiefly connected with marine action, as described in Chapter III: e.g. cliff coasts like those of the Pays de Caux and dune coasts.

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Chapter V

COASTAL EVOLUTION

Classic concepts of coastal evolution. Coastal evolution results from marine agents working upon landforms which initially owe nothing to their action; it leads to the development of a sequence of new forms. The trend of this evolution caused by the action of the sea must be analysed. It has been universally accepted, since the works of the early authorities on coastal morphology, that the sea tends to regularize shorelines. Except for coasts bordering extensive unglaciated lowlands, all initial types of coasts are irregular: rias, fjords, various morainic forms, and irregularities due to structure all create numerous indentations in the shoreline. The work of the sea tends to reduce these indentations both by eroding the headlands and by filling the bays. In effect the refraction of the swell by the shoals (p. 16) causes the wave crests to converge on the headlands and to diverge in the bays, at least in the common example where the underwater relief offshore continues the relief of the adjacent land mass. Erosion of the headlands results in cliff formation by processes already discussed. The filling up of bays takes place in two ways; partly by direct deposition in the bays, and partly through the formation of spits, or bay-mouth bars, which accelerate sedimentation behind them by reducing or preventing the movement of water therein. Once the coast is smoothed out, it retreats inland parallel to itself: depositional forms are in fact driven back with the rocky headlands to which they are attached, and after a long time, the whole coast can recede to the line of the old bay-heads and even beyond. This retreat is accompanied by the development of the so-called abrasion platform, which should be termed the marine erosion platform since it is not due simply to mechanical action. There has been considerable discussion whether the coast can be eroded back indefinitely. In fact, the feebleness of marine erosion at and below low tide level (Berthois) leads one to believe that it cannot, unless subsidence occurs as well. In marine erosion and shoreline development a balance tends

to develop, as Baulig says, between the forces of erosion and the resistance of the rocks. Modifications continue to occur but ever more slowly, for, as erosion proceeds, the efficacy of the forces decreases.

Verification of the classic concepts. It would not be difficult to verify the accuracy of these concepts if sea-level had been stable for a long time: we could then ignore tectonically unstable regions, and concentrate on stable areas. But this is unfortunately not practicable. Apart from minor oscillations, such as the very slight transgression which now results from the reduction of the volume of the world's glaciers and ice-sheets, the extensive Flandrian transgression is too recent: its final phase, the Dunkirkian, probably took place only about two thousand years ago, and it is generally agreed that in the Neolithic period the transgression was still in operation since certain Megalithic monuments in such stable regions as the Morbihan are now several metres below sea-level. All the coasts of the world, therefore, were, after the Flandrian transgression, in a stage of extreme youth resulting from the submergence of subaerial relief. The Flandrian transgression probably restored within a few metres the pre-Würm sea-level, i.e. the Low Monastirian level. But the sea stood at this level only for a limited period, probably longer than it has stood at its present level but not long enough for marine forces to have had sufficient time to produce mature forms on all the shores of the world.

It seems, therefore, that the only means of verifying the classical concepts, or, to be more precise, the reality of the tendency to regularization which they postulate, would be to consider only those coasts which evolve rapidly, especially morainic coasts: as we have seen (p. 70), the retreat of such cliffs takes place so quickly that a man's life is long enough for changes of several dozen metres to be observed. Such recession gives the sea enormous amounts of material which may later be built up into various constructional forms. The morainic coasts of north Europe and New England have been the object of excellent and very detailed studies by Johnson in America and Schou in Denmark, who confirm the accuracy of the classical concepts. Three examples will illustrate this.

(*a*) In the bay of Boston, there is an east-facing coast with an almost regular outline in the neighbourhood of Winthrop. It includes beaches and curved spits, and drumlins or parts of drumlins to which the constructional forms are tied. The position of former drumlins on the shore, or a little farther seaward, is known, thanks to the presence of residual boulders. It is therefore possible to reconstruct the shoreline of the

period before these drumlins had been planed down by the sea; that shoreline was more irregular than the present one (Fig 26 A) In this case, there has been erosion of the headlands, smoothing-out of the bays by the formation of beaches or spits, and a retreat of the whole as the erosion of the drumlins continues

(*b*) At Martha's Vineyard, near Cape Cod, there is a long, straight, south-facing beach bordering old proglacial outwash forms (Fig 26 B)

A Boston Harbour, New England (after Johnson), B Martha's Vineyard, New England (after Johnson) C South Zealand, Denmark (after Schou)

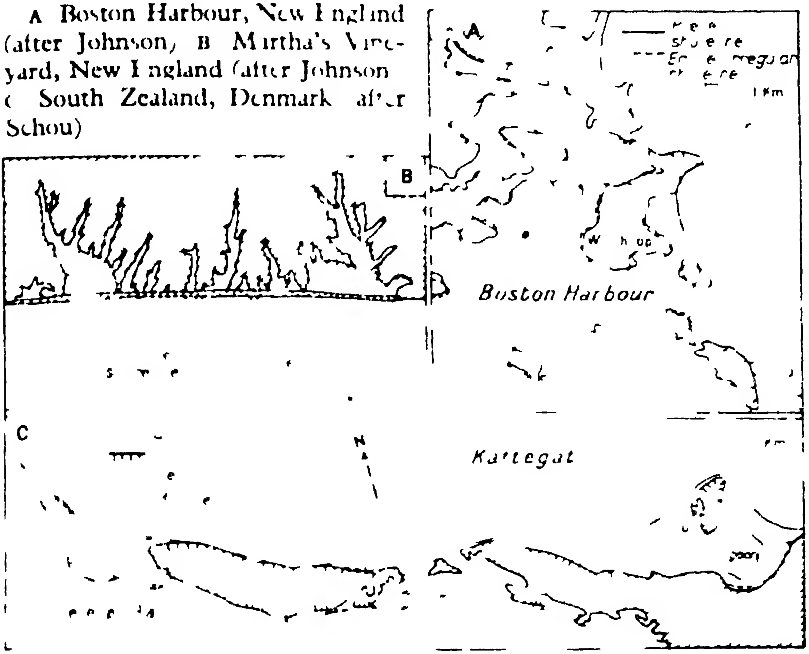


FIG 26 REGULARIZATION OF MORAINIC COASTS

Such forms are slightly convex, and consequently, the initial shore, as has been said on p 161, possessed wide convex lobes, which are shown up on the map by the flooded radial valleys cut into them An examination of the map shows that the outermost parts of the central and eastern cones have been cut off by the sea Regularization has again been achieved by the formation of bars across the bays and by erosion of the headlands as at Winthrop, a general retreat has occurred

(*c*) The north-west coast of the island of Zealand in Denmark is shown in Fig 26 C It consists of low areas of marine sedimentation, and higher cliffs cut in glacial deposits These cliffs are partly live and partly dead The cliffs are found in the projecting parts of the morainic

hills where they have been attacked: thus, the hills initially formed headlands more pronounced than those of today. In the most sheltered parts there are no cliffs at all because deposition took place there first. In the intermediate parts, there are dead cliffs (right centre of the figure) in places where deposition has followed an early phase of cliff erosion. Here again there has clearly been a progressive regularization, which will be completed by the erosion of the former island at present connected to the north coast by a tombolo, and by a parallel retreat of the shoreline as at Martha's Vineyard.

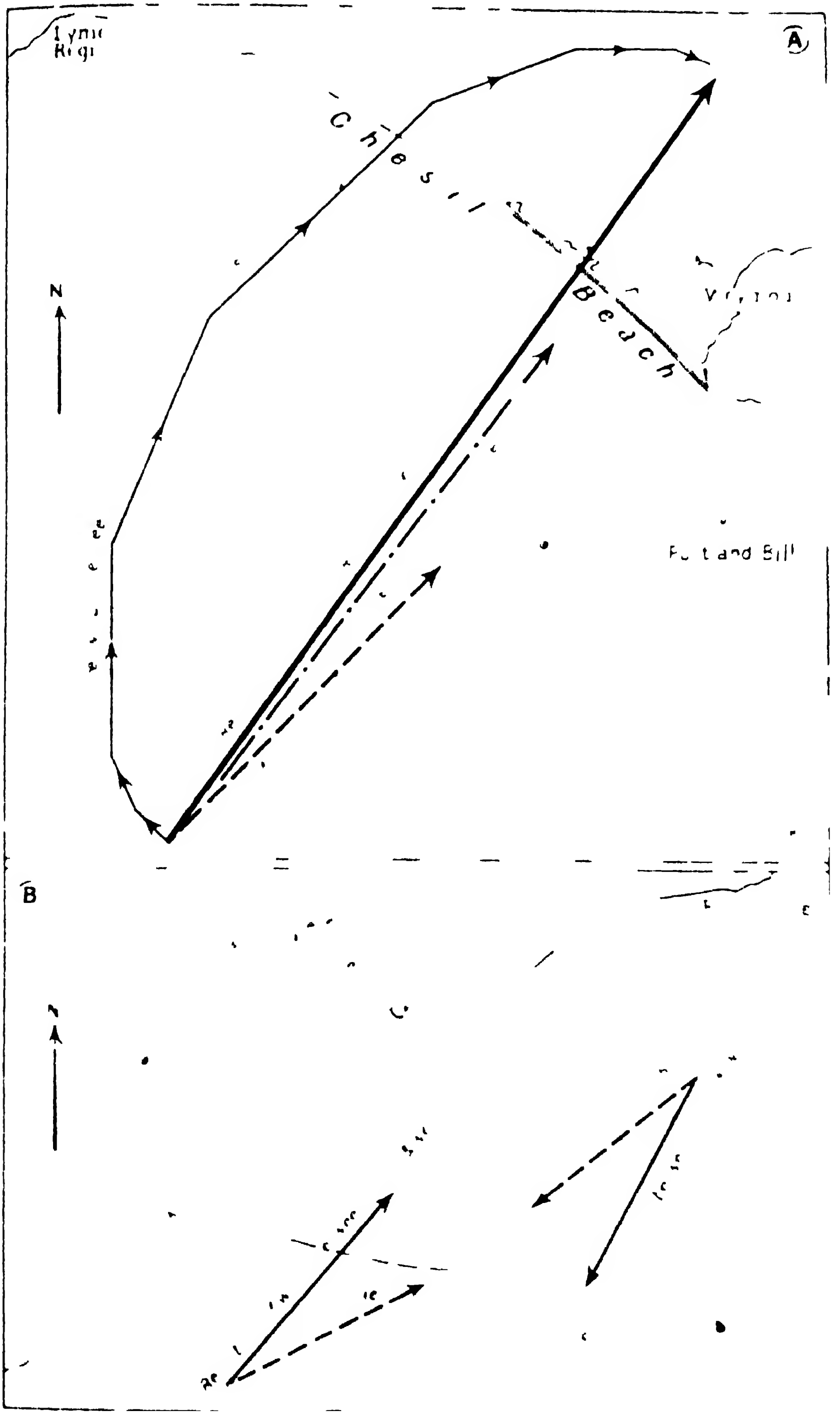
It must be clearly emphasized that these examples in morainic regions are much better illustrations than the ria areas in hard rock. Even if there is regularization of the latter as a result of the formation of spits parallel to the general trend of the coast, one seldom sees clear proof of any significant retreat of the headlands. On such a coast only one of the two processes of the classical theory is realized. It seems that the classical theory has been evolved from the study of morainic coasts. These coasts are found in regions where the Americans, the English, and the Germans had every opportunity to work, and it is not just by chance that the theory was first developed in these countries, especially in America.

The orientation of coasts. Although the classical theory contains the idea of the general trend towards regularization, it does not include any ideas on the orientation of the regularized coast. This gap has been made good by Lewis and Schou. The former has formulated the following rule: *beaches tend to orientate themselves perpendicular to the dominant waves.* Schou amplified this idea by a more precise consideration of the maximum fetch and the resultant of strong winds, exceeding Beaufort force 4 (18 m.p.h.). The resultant of the winds is calculated in the following way: the frequencies for each direction are multiplied by the Beaufort force (or better by the actual velocities: Williams even thinks *(in litt.)* that it is better to multiply the frequencies by the cube of the velocities). The totals are added vectorially and the resultant is the straight line joining the first and last points (Fig. 27 A). From this Schou formulates the following laws:

(*a*) When the fetch is equal in every direction, the trend of the

FIG. 27 CHESIL BEACH AND DUNGENESS IN RELATION TO WIND AND FETCH

Wind data kindly supplied by the British Meteorological Office. The frequencies have been multiplied by the cube of the velocity.


A
Chesil Beach
N
B

coast is at right angles to the resultant of the wind, both on advancing and retreating coasts, because it is the direction causing the minimum of longshore drift, which really enables the orientation to be maintained.

(*b*) When the maximum fetch and the resultant of the wind are coincident, conditions are similar.

(*c*) When the maximum fetch and the resultant of the wind do not coincide, the coast aligns itself perpendicular to a line between the direction of maximum fetch and the direction of the wind. The exact direction depends upon the relative importance of these two factors, which may vary.

(*d*) Finally, the coast may sometimes be orientated not perpendicular but parallel to the resultant of the wind. The rule holds for banner-banks (p. 91) and forms having only one point of attachment. The other laws apply particularly to coasts of erosion and constructional forms attached to the coasts at both ends. Nevertheless, for reasons which are obscure, certain forms free or unattached at one end sometimes obey the first three laws.

It will be appreciated that these laws only indicate tendencies, and that initial forms or structures, in a very great number of cases, control the present orientation of the shores.

The laws of Lewis and Schou can be illustrated well in the Baltic and on the south and west coasts of Great Britain. In particular Chesil beach near Portland illustrates Lewis's law and, in more detail, the effect of Schou's third law: it is orientated perpendicularly to a direction between that of maximum fetch and that of the resultant of the winds at the nearest recording station, which is Calshot at the entrance to Southampton Water (Fig. 27 A). Chesil beach may, therefore, be considered as being in equilibrium. The large beaches in the north of Cardigan Bay at Aberdaron, Porth Neigwl, Morfa Harlech, and Morfa Dyffryn obey the same law: they face a direction intermediate between that of maximum fetch through the entrance to St George's Channel and the resultant of the winds at Holyhead. On the other hand, the small beaches near Pwllheli still remain under the influence of initial structures. Finally, the two sides of Dungeness tend, as has been said on p. 91, to be orientated perpendicularly to directions between the maximum fetches (through the entrance of the English Channel and from the Dutch coast) and the dominant winds, south-west on the south side and north-east on the east side. Fig. 27 B shows that the evolution, witnessed by the pattern of shingle ridges (Fig. 8 B and C), is not finished. It may be

predicted that the south coast of the ness will finally run north-west–south-east, that the ness will move further out into the Channel in which direction the fetch is very short and the winds light and variable; and that at the same time it will migrate towards the north-east, because the forces concerned are unequal. Beniguet (p. 90 and Fig. 7 D) illustrates Schou's fourth law since it is parallel to both the fetch and the direction of the wind.

Dungeness, however, shows the necessity for modifying Schou's law in certain cases: as two wind directions are significant in its formation, we must make separate vectorial diagrams for two sectors, south to west and north to east. It is also necessary to eliminate offshore winds not only at Dungeness, but in every case except where only a small island is concerned. These winds have been eliminated in the diagram of Chesil beach (Fig. 27 A). After an extremely long evolution, the shore may so change that winds from a certain direction, which were offshore winds in the early phases, may become onshore winds, but such considerations are entirely theoretical as changes of base-level usually intervene long before such a condition is realized.

Schou's theory is based on the supposition that the effective waves are caused by local winds. This supposition is certainly valid in enclosed seas. For coasts open to the ocean, it is no longer necessarily true, for swell of distant origin may predominate. The direction of such swell may not conform to the resultant of the local winds, as has been shown in the case of the Landes of Gascony by Guilcher, Godard, and Visseaux, and for the whole of the Atlantic coast of Africa by Jessen and Guilcher. In such cases the swell must be analysed, but data on the swell are often more difficult to obtain than local wind statistics.

Finally, in examples where the swell is caused by local winds, Berthois considered that, when anemometers measuring the run of the wind are used, there was something to be said for replacing the product of wind frequency and velocity by the number of kilometres of wind passing in each direction. In this way one would obtain a more exact measure of the relative importance of different winds. This method involves difficulties, one of which is due to the fact that such anemometers have often been only recently set up, so that only a short period record is so far available. They also sum up all the winds, even those equal to or less than Beaufort force 4 which are eliminated by Schou, probably rightly, as ineffective. Finally, it is difficult to reconcile this method with the fact that waves do not increase only in arithmetical progression with the velocity of the wind, for one can express their effect by the multiplication

of the frequency by the cube of the velocity as Williams proposes. But Berthois' method has proved excellent for the study of short-term effects of local storms on beaches.

Irregularization of coasts by differential marine erosion. Although morainic coasts show what happens in the final stages of evolution, certain coasts of hard rocks and varied structure show what may happen in the early or intermediate stages.

The adoption of the classical theory led to the rejection as heretical of an older theory according to which the sea had eroded the bays. Much of this theory, such as the view that rias are the work of tidal currents, must be abandoned.

Yet, even the most orthodox authors admit at times that the sea may make the coast irregular in the initial stages of evolution. Gulliver (*Shoreline Topography*, p. 173) states that 'waves will attack softer rock more rapidly than its more resistant neighbour. A promontory of hard rock may thus be formed where the less resistant rock on either side has been eroded by the sea.' He goes on to say that coasts will be regularized if they remain stable for a sufficient length of time. Davis (*Erklärende Beschreibung der Landformen*, p. 502) considers that[1] 'if a coast including materials of great variety is attacked by waves of exceptional force, the resulting coast line will be very indented before beaches are built to simplify it. In these conditions, old bays are widened . . .; soft rocks favour a jagged coastline . . .; it is only later that the regularization of the coast begins.' And Cotton states (*New Zealand Geographer*, Oct. 1951, p. 110): 'Across a terrain of heterogeneous rock the marine erosional process does not always develop a straight line of cliffs; on the contrary, as a result of adjustment to structure, the outline of the shore when it has reached maturity is commonly sinuous and follows the pattern of the terrain.' He returns to this (ibid., April 1952, p. 56) and again admits that this adjustment to lithology lasts a very long time (in this, therefore, he differs from Davis) saying: 'This is perpetuated as long as the coast continues to be cut back across terrain with similar structure.' Finally Shepard (*Submarine Geology*, p. 75) speaks of 'sea cliffs made irregular by wave erosion . . . . with small bays in contrast to . . . drowned river valleys'.

These passages have generally remained unnoticed in France, where writers seldom admit selective marine erosion except in connexion with the formation of caves and small fissures. These views may be summed

[1] Professor H. Baulig kindly drew the author's attention to this passage in Davis's work.

up as follows: the sea is capable of differential erosion in certain conditions, at any rate in the initial stages and even, according to Cotton, in maturity. This holds not only for minor details but also for the general erosion of the coast.

This seems to be a perfectly sound idea, at least for the initial stages of evolution. Observations made in Brittany, Cornwall, and Wales completely confirm it.

These areas offer good examples of contraposed coasts (pp. 171–2). Most authors, especially Johnson and Clapp, agree that on this type of coast the sea begins by clearing away the unconsolidated deposits from the depressions they fill, before eroding the solid rock below. Thus the coast is made irregular by the exhumation of the underlying relief from under its smooth cover of drift. That is exactly what happens in the bays of Aberdaron and Porth Neigwl in the Lleyn peninsula and in Rhossili Bay in the Gower peninsula (pl. VIII B). The sea causes rapid recession in the glacial deposits at the heads of the bays where the cliffs have typically slumped forms, although it has had as yet virtually no effect on the [illegible]ing headlands formed of Paleozoic crystalline or volcanic rocks. At Worms Head, which shuts in Rhossili Bay to the south, the absence of recession since the Flandrian is proved by the presence of several old beaches antedating the last glaciation (Fig. 4 J). The coast of northern Brittany has developed in a similar way. The confused arrangement of granite rocks which often characterizes this coast is explained simply by the exhumation of an underlying surface. These rocks occur buried under loess, solifluxion deposits, or products of weathering, and the sea merely exhumes them. Around Saint-Briac, Paimpol, Plougrescant, Ploumanac'h, Plouescat, Kerlouan, and in other places, all stages of this progressive irregularization, from the simple hollowing out of coves in loess between granite headlands to the complete isolation of rocks on the shore after the total removal of the loess, can be seen (Fig. 25 A). In Cornwall, Penzance Bay is in process of being extended in soft rocks, although there is little erosion on the Land's End granite massif on its side, as the Lower Monastirian beaches along it show. In Jersey the bays of Saint-Ouen and Bonne-Nuit, for example, are being cut into loess between uneroded rock headlands. All these coasts, like the comparable ones in New England, are in the youthful stage of evolution.

Apart from contraposed coasts, examples of large-scale differential marine erosion leading to very indented coastlines may be found. South Brittany provides two good examples, the bay of Audierne, and the

embayment between Gavre, near Lorient, and Quiberon. The bay of Audierne (Fig. 28 A) is formed in deeply weathered mica-schists enclosed within the crystalline headlands of Cape Sizun, which ends in Raz point, to the north and of Cape Caval (Penmarc'h) to the south. Neither on Cape Sizun nor on Cape Caval does the solid rock show signs of any marked erosion since the Flandrian transgression, because numerous fragments of the Lower Monasturian beach are preserved

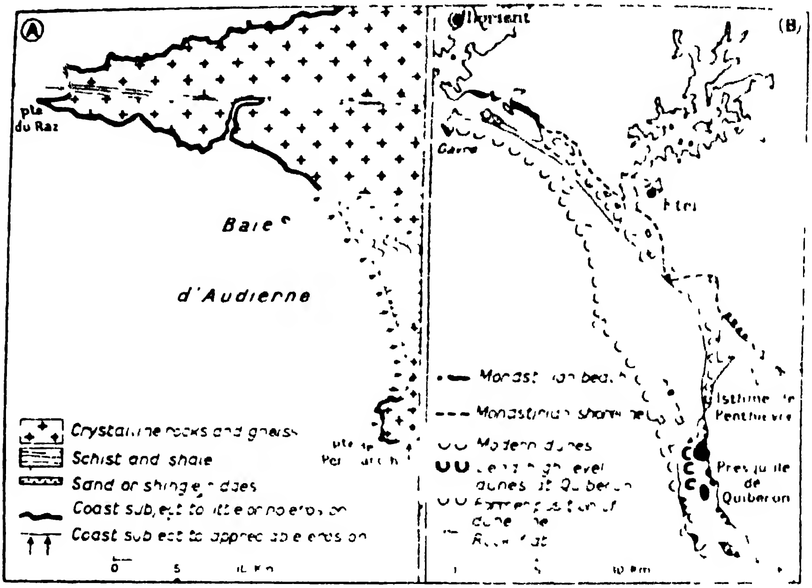


FIG. 28 SHORELINE EVOLVING BY RECESSION IN THE EMBAYMENTS IN SOUTH BRITTANY

A. The bay of Audierne. B. Between Gavre and Quiberon.

below the cliffs as at Worms Head. In the middle of the bay, on the other hand, erosion is rapid: at Pehors 30 m. have disappeared in fifty years, and at Lessunus some 25 m. in twenty years. Although erosion has not been sufficient to destroy the Lower Monasturian beach which occurs on this low coast, it has certainly tended to accentuate the concavity of the bay. Between Gavre and Quiberon (Fig. 28 B) there is similarly good reason for thinking that the bay was formerly not so marked: the line of dunes which borders this shore has left evidence of its former more westerly position in the shape of dead dunes on the west coast of Quiberon. Coastal evolution is tending to separate Quiberon from the

mainland by breaching the tombolo of Penthièvre. The construction of a sea-wall has prevented a breach being formed as a result of wave-action in Quiberon Bay.

The bay of Audierne and the beach from Gavre to Quiberon are examples of another important fact already mentioned on p. 86: a *beach coast is not necessarily a coast of progradation*, i.e. one being built out seawards. In fact these receding shores are characterized throughout by sand and shingle formations. Similarly the only parts of the Ile de Sein, where a noticeable recession is recorded, are bordered by shingle beaches. Such conditions are readily understood. The sea cannot do much by itself, but needs to be armed with pebbles to perform mechanical erosion. The pebbles may cause erosion of the rock platform in the intertidal zone in front of the main beach, or erosion of the cliffs, if these are present, by being flung against them. On low coasts, coastal ridges may move quickly inland as can be seen when beaches move inland over coastal marshes where they compress the peat beneath them (p. 109). Progradation occurs only when the waves are overloaded with material, either in amount or calibre.

It is therefore apparent that differential marine erosion occurs on a large scale as well as in detail on coasts other than contraposed shorelines where it always happens. It seems to occur when the lithological differences are great, as in the bay of Audierne and on contraposed coasts, or when bays are still widely open to strong marine action, as in the bays of Audierne and Aberdaron, and between Gavre and Quiberon. But to what stage of coastal evolution can these processes continue? The Breton and Welsh examples cannot provide the answer, since evolution is in such an early stage that the coasts have not been eroded back to their Lower Monastirian position. It is true that the cliffs of New Zealand recede rapidly, as Cotton says, then this differential erosion may continue for quite a long time. Because Cotton believes in a rapid recession of these cliffs, he thinks, unlike Gulliver and Davis, that differential erosion may affect even mature coasts. It is certain, however, that bays cannot be cut indefinitely landwards while the headlands remain uneroded, for in bays extending some way inland friction would considerably reduce the power of the waves.

We may conclude that, although coastal evolution leads to the regularization of the coast and to the reorientation of the coast to face the dominant swell, it is no less certain that, where rocks of very varied resistance occur, the sea often makes the coast more irregular at first, even though it may still tend to orient it in the same way. There is still

uncertainty as to what happens on coasts of varied structure in an advanced stage of evolution.

Evolution may be interrupted by several factors. The closing of a strait may lead to changes in longshore drift and deposition at the foot of a cliff which then becomes dead. The formation of a new strait may have the opposite effect. Intermittent uplift may cause the formation of a number of cliffs, as in the Ventura region in California. The recession of such a shore may, if erosion is rapid enough, cause all these cliffs to combine into a single series. Individual examples such as these may be readily deduced from general principles.

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PART TWO

An Outline of Submarine Geomorphology

Chapter VI

GENERAL REMARKS

Methods of Research. The difficulties and slowness of submarine morphological research have already been mentioned in the Introduction. However, even apart from direct exploration which will certainly take place in the near future, present methods of investigation have been greatly improved in the last few decades, so that modern studies of the sea-floor have reached a level never envisaged at the beginning of this century. This research has not been purely speculative: oil companies are very interested in the subject and one has only to read publications, such as the *Bulletin of the Geological Society of America* and the *Bulletin of the American Association of Petroleum Geologists* to realize the importance attached to deep-sea research.

Such work is very costly, and individual work, which is still common in ordinary geomorphology, has no longer any place here. Team work with a research ship, often attached to an institute, is the only possible means of doing work of real value. The institution plans the operations at sea and works up the results: it is absolutely indispensable, if the field-work includes both sedimentology and morphology. These two sciences are even more closely associated than in research on land. One might go so far as to say that geology as a whole and morphology are practically inseparable. Examples of well-equipped institutes are found at La Jolla, California, Woods Hole, Massachusetts, Hamburg, and Wormley in Britain. The French institutes are essentially concerned with marine biology or hydrology, except for the Sorbonne and Villefranche laboratories, under the direction of Bourcart, and the La Rochelle institute, which comes under the C.R.E.O., although some, such as the Institut des Pêches Maritimes, also study the relief of the sea-bed.

Nearly all the principal countries possess oceanographic research ships: most of them study the sea-bottom, hydrology and biology at the same time. The *Pourquoi Pas?* disappeared in a storm off the coast

of Iceland, but France now has the *Ingénieur-Elie-Monnier*, the *Calypso*, and the *Président-Théodore-Tissier*. The first and second are chiefly concerned with underwater research; the third is used in investigating ways of increasing the catch of fish. The *Elie-Monnier* and *Calypso* possess good equipment for morphological research. The United States has the *Horizon* (Scripps Institution of La Jolla), the *Atlantis* (Woods Hole Institution), which operate respectively in the Pacific and Atlantic, and many others. Great Britain has principally the *Discovery II* and the *Challenger*. Most of these ships are small; they vary usually from 140 to 500 tons, but the *Discovery II* is larger. A shallow-draught vessel is essential for studying the coral seas. Among the ships which have made considerable contributions to our subject, we should mention the first *Challenger*, an English ship, the German *Meteor*, and the Dutch *Snellius*.

For a long time these ships were equipped for deep soundings only with lead lines, measurement being reckoned by the number of turns on the cable drum. A sounding of 4,000–5,000 m took several hours, so that the slowness of the advance of knowledge is easy to understand. About 1919, only the shallow seas near the countries of European civilization and the routes of underwater cables were known with any accuracy. In 1919, the French hydrographic engineer Marti introduced a revolutionary method, echo-sounding. The use of supersonic waves at a later date was another great advance, as such waves are directional. There is less chance with such waves of receiving echoes from objects to one side rather than from underneath the ship. Modern apparatus allows continuous soundings to be made while the ship is in motion. The soundings are automatically recorded on a chart, thus giving a continuous trace of the depths. The results are very satisfactory, except when the sea is rough or when the ship's hull is covered with algae. On an expedition, on which it may be impracticable to scrape the ship's bottom, it may be necessary to steam into a river to kill the algae. From a close network of soundings a contoured chart may be drawn up, but sharp isolated peaks and small depressions may be missed, because such features cannot be seen as they can on land. The positions of the ship must be carefully recorded in order to plot the lines of soundings on the chart. They are obtained by different methods; direct fixes near the coast; astronomical fixes, radioacoustic ranging, radar navigating devices out at sea. In spite of the advances brought about by supersonic soundings, the gaps are still enormous, especially in the southern hemisphere and in the Arctic basin.

The nature of the bottom, provided that it is not rocky, may be ascertained by using lead lines. Detail recorded on navigation charts has been obtained in this way. But this is a crude method and much more effective apparatus is now available for the purpose. Bottom-sampling devices are carried by all oceanographic ships; they consist essentially of corers, long tubes which sink into the sea-bottom, unless it is rocky, in which case the corers are liable to be broken. Some, such as the Piggot cannon, are forced into the sea-bottom by the use of explosive. Others, such as Kullenberg's piston core-sampler, rely on their own weight. Kullenberg's core-sampler can produce cores 20 m. long in soft deposits: it thus gives sections of Quaternary and even Tertiary deposits, which have provided the best information about Quaternary climatic changes (p. 258). But this gear is very heavy (1,500 kg.), difficult to handle, and the interpretation of each core demands months of work in a laboratory. The Stetson corer, easier to handle, extracts cores about 10 m. long

Samples of consolidated rock may be collected with the Charcot dredge or similar apparatus. These dredges also bring up sediments, such as pebbles, which are too coarse to enter the piston corers. Both types of apparatus have to be used, therefore

The sea-bottom may be photographed either by apparatus carried by divers in shallow waters see below or, at greater depths, by cameras suspended from a cable Some additional source of light is necessary beyond depths of about 10 m Submarine photography is one of the most delicate techniques in oceanography, and successful pictures are not numerous, although of great interest. Good results have been obtained from depths of 5,000 m and more Various underwater films, usually concerned with marine life, have been sho in cinemas, notably those of Commandant Cousteau.

Geophysical methods may be used for obtaining structural data to considerable depths below the sea-floor The methods are the same as those used in prospecting for oil on land. Either seismic soundings may be made, or magnetic data, obtained by a flight over the region concerned, may be used. In 1952–3 the American *Capricorn* expedition used a marine magnetometer, towed by a boat. The first method, the seismic method, makes use of the different velocities of sound waves in different rocks and the reflection of such waves from surfaces of discontinuity. The waves, produced by exploding small charges, make it possible to distinguish between unconsolidated sediments, consolidated sediments and the underlying crystalline masses. The volume of data

is becoming large for certain regions, especially for the areas off the east and south coasts of the United States, and Caribbean Sea. These methods have also been used in the study of Bikini atoll (p. 132). It is, therefore, possible to evaluate submarine oil resources, and already oil wells are being drilled in shallow water off the north coast of the Gulf of Mexico.

But none of these methods allow one to *see* the bottom of the sea The problem of direct observation has now been solved for depths up to 70 m. by the Cousteau-Gagnan diving apparatus, which uses compressed air and was first manufactured in 1943. It is an improved form of the Rouquayrol-Denayrouse and Le Prieur apparatus This apparatus is much less dangerous than the ordinary diving suits used by specialists, as these are death-traps if the pressure balance is not accurately maintained. The new apparatus offers the great advantage of entire freedom of movement, since the diver is not attached to the boat by a cable. His movements can be followed by the air bubbles released whenever he breathes. Most divers can be quickly trained to descend to a depth of 25–30 m Beyond 40 m there are risks of nitrogen giddiness, which become very serious below 70 m, so that only very skilled divers can operate at this depth This diving apparatus, invented in France, has unfortunately not yet been widely used by morphologists, although a biologist, P Drach, has clearly realized its value, and submarine archaeology has also been advanced by its use

It is possible, but much more difficult, to dive to depths exceeding 70 m. This can be done either by means of a rigid, articulated diving suit supplied by a pump, like those used by the Italian company S.O.R.I.M.A. for the recovery of wrecks, or with self-contained Brown diving suits of American origin in which one breathes below a depth of 60 m. a mixture of helium and oxygen, which is unsuitable for breathing at ordinary atmospheric pressure. This apparatus allowed a depth of 163 m. to be reached in 1948; but it cannot be used by amateurs such as the morphologist.

Another way of seing the bottom is to go down in some sort of globe equipped with portholes. Such apparatus can reach greater depths than those reached with diving suits, and may well become more popular The first globes to be used were Beebe's *Bathysphere* and Barton's *Benthoscope*. The latter reached a depth of 1,375 m. in 1949, and 1,100 m. in 1952. Both have the same limitations as diving suits supplied by pumps: i.e. they are not self-contained. An independent and self-propelled bathyscaphe designed by Cosyns and Piccard, did not

work satisfactorily in its trials at Dakar, but two improved bathyscaphes were much more successful in the summer of 1953: Piccard reached a depth of more than 3,000 m. near Naples, while a bathyscaphe belonging to the French Navy reached below 2,000 m. near Toulon and below 4,000 m. off Dakar in January 1954. Such apparatus means enormous progress in underwater research.

Finally, underwater television is possible by means of Wexler's electronic periscope (1948). This has already been used successfully and should be capable of further improvement.

This apparatus and the bathyscaphe seem to offer the best methods of making a visual study of the sea-bed. The portable diving suit of Cousteau-Gagnan is a really practical apparatus, simple and easy to manage down to a limited depth. The technique of its use in underwater geology and mapping by divers has been recently established and described by a team of American geologists.

Marine charts may be grouped in three categories.

(*a*) Fishing charts are the least important because they only cover recognized trawling grounds and then only incompletely. They show depths, contours, and indications of the nature of the bottom. They are of value for the morphological study of the areas they cover.

(*b*) Oceanographic charts, especially the General Bathymetric Chart of the Oceans on a scale of 1 10,000,000, edited and kept up to date by the Bureau Hydrographique International de Monaco. This chart is layer-coloured, actual depths being given in metres. It covers all the oceans, but its value varies with the density of the network of the soundings. As many soundings as possible are plotted; this is a relatively easy task since place-names are few. In certain countries, especially America, larger scale charts of the same type covering more restricted areas have been published. They cover fairly well-known areas, such as the coasts of the United States and the Western Pacific, and are very useful.

(*c*) Navigation charts have a definite purpose. They are the successors of the portolan charts of the fifteenth and sixteenth centuries, and the 'Neptunes' of the seventeenth and eighteenth. Great progress was achieved in France, under the direction of Beautemps-Beaupré, at the beginning of the nineteenth century. The hydrographic offices of all the maritime nations publish such charts; they make use of the data published by other countries for regions where they themselves have not made any surveys. In England navigation charts for any region in the world may be bought, but for certain areas more detailed

and precise sheets are available from Paris or Washington. These charts are the best documents for coastal areas, since they are constantly being brought up to date, and their scales sometimes exceed 1/15,000. But the scales are very variable, and it is not usually possible to cover a large area on one fixed scale. These charts are drawn on Mercator's projection, which is suitable for sailors since bearings are correct.

Unfortunately, navigation charts do not all use the same datum. It is hardly an exaggeration to say that there are as many different datums as there are countries. In France, the level of the very lowest tides is used. This, like all other levels except mean sea-level, varies with the local tidal range. The datum on a chart of the coast of Provence, for example, is more than 5 m. above that of a chart of the approaches to St Malo. But the use of this datum has great practical advantages for navigation. The British Admiralty has chosen the mean level of low water springs: the U.S. Coast Guard and Geodetic Survey has taken mean low water for the Atlantic coast and the mean low water of the principal diurnal tides for the Pacific coast. French charts based on foreign charts keep the datum of the original charts and carry a notice to that effect. Those overlapping several countries, as in the Straits of Dover, have more than one datum. The soundings are in fathoms on British and American charts, and in metres on French charts, where tenths of a metre are given down to depths of 10 m.; as many as possible are shown. A high degree of accuracy is obtained for frequented seas and in depths down to 20 m. Below this the soundings become much less frequent on most charts. Not all charts show contours of the sea-bottom, although they may be readily interpolated from the soundings. Their limitations should be realized.

The regions of the sea-floor. There are two ways of studying the relief of the sea-floor: either through a consideration of the relief of the geographical regions of the sea-bottom or through a systematic division according to depth and distance from the coast. In this work we must adopt the second point of view, since we are concerned only with general geography. Certain fundamental features may be stressed. The Pacific and adjacent seas cover about 50 per cent. of the total sea area of 361 million sq. km.; the Atlantic covers 106 million sq. km. and the Indian Ocean 75 million sq. km. The southern hemisphere is mainly oceanic since between the Equator and 70° South less than 25 per cent. of the area is land, whereas land exceeds 50 per cent. between 45 and 70° in the northern hemisphere. The South Pole is,

however, occupied by a continent, while at the North Pole there is a deep basin. The distribution of land and sea is therefore roughly antipodal.

The nature of the general distribution of depth is fairly well known and further soundings will not greatly alter our knowledge of it. It may

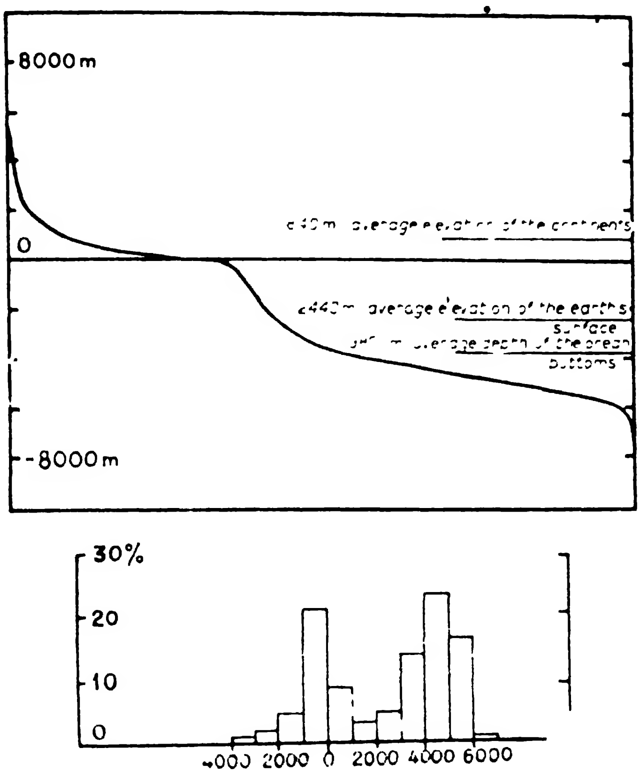

FIG. 29 DISTRIBUTION OF LAND AND SEA (AFTER SVERDRUP, JOHNSON, AND FLEMING)

Hypsographic curve and percentages of total area between successive 1,000-m. contours.

be expressed by a hypsographic curve (Fig. 29), which shows two dominant areas: one between 0 and 1,000 m. on the continents, with an average altitude of 840 m., and the other between 3,000 and 6,000 m. in the oceans with a mean depth of 3,800 m. The oceans are therefore generally much deeper than the continents are high. The maximum depths exceed the greatest heights. The deepest sounding up to 1956

was 5,940 fathoms or 10,863 metres in the Marianas trench. The percentage of the total area at different depths is as follows:

0– 200 m.	7·6%	
200–1,000 m.	4·3%	8·5%
1,000–2,000 m.	4·2%	
2,000–6,000 m.	82·7% (of which 33% lies between 4,000 and 5,000 m.)	
Below 6,000 m.	1·2%	

This table reflects certain fundamental elements in submarine relief, the terms for which have been fixed by the International Committee on the Nomenclature of Ocean Bottom Features (Brussels and Monaco, 1951–2):

(*a*) From 0 to 100 fathoms or 200 m. is the *continental shelf*, which some authorities consider to extend to 500 m.

(*b*) From 200 to 2,000 m. is the *continental slope*, which is often cut by *submarine canyons*; the table above shows that this zone is steeper than the continental shelf. These two zones are grouped by some authorities as the *continental terrace* or *continental margin*; this seems to be reasonable, since their origins appear to be connected. We will therefore study them together in Chapter VII.

(*c*) From 2,000 to 6,000 m. are the *general ocean depths* which have no universally recognized name, but which include *troughs* (broad, elongated depressions with gently sloping sides), *basins* (broad, circular, or oval depressions, in which the parts below 6,000 m. may be called *deeps*), *rises* (broad, long elongations with gentle slopes), and *ridges* (long, narrow elevations with fairly steep slopes).

(*d*) Below 6,000 m. there are the deeps, already mentioned, and *trenches*, which are narrow and steep sided. The area they occupy is very small compared with that occupied by the general ocean depths.

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Chapter VII

THE CONTINENTAL MARGIN

A. EXTENT AND RELIEF OF THE CONTINENTAL SHELF

The continental shelf is usually wide off low-lying continents and narrow or absent off mountainous regions. This relationship between continental and submarine relief was realized very early in the study of the oceans.

The continental shelf is wide in north-west Europe, off the north coast of the U.S.S.R., in the China Sea, the Arafura Sea, the Gulf of Lions, the Gulf of Gabes, the north Adriatic, the east coast of North America. It is narrow in front of the west coast of North America, and the north coast of New Guinea; very narrow or absent off the Provençal coast, where, for example, the Ile du Levant is separated from Cap Lardier (Maures) by depths of more than 1,000 m. and the submarine relief is as marked as that of the Maures Massif. The average width of the continental shelf is, according to Shepard, 42 miles.

The continental shelf meets the continental slope at a slight angle, but this break of slope is not always at the same depth and it is often difficult to locate precisely. According to Shepard it is found usually between 70 and 100 fathoms, the average being 72 fathoms. The mean slope of the continental shelf is 0° 07'; it is a little steeper near the land than in its seaward parts. The average depth of the flattest parts of the continental shelf is 35 fathoms.

However, P. Birot points out that there is often a continental shelf, 20 km. or so in width, off mountainous coasts, thus breaking the continuity between the terrestrial and submarine relief. Off Galicia, for example, where the land reaches 400–500 m. close to the sea, depths of 200 m. are reached only at a distance of 20–30 km. from the shore.

In some places areas with gentle slopes are found below 200 m. J. Bourcart thought he had found a series of such surfaces, which became progressively steeper seawards, off Morocco, but their existence

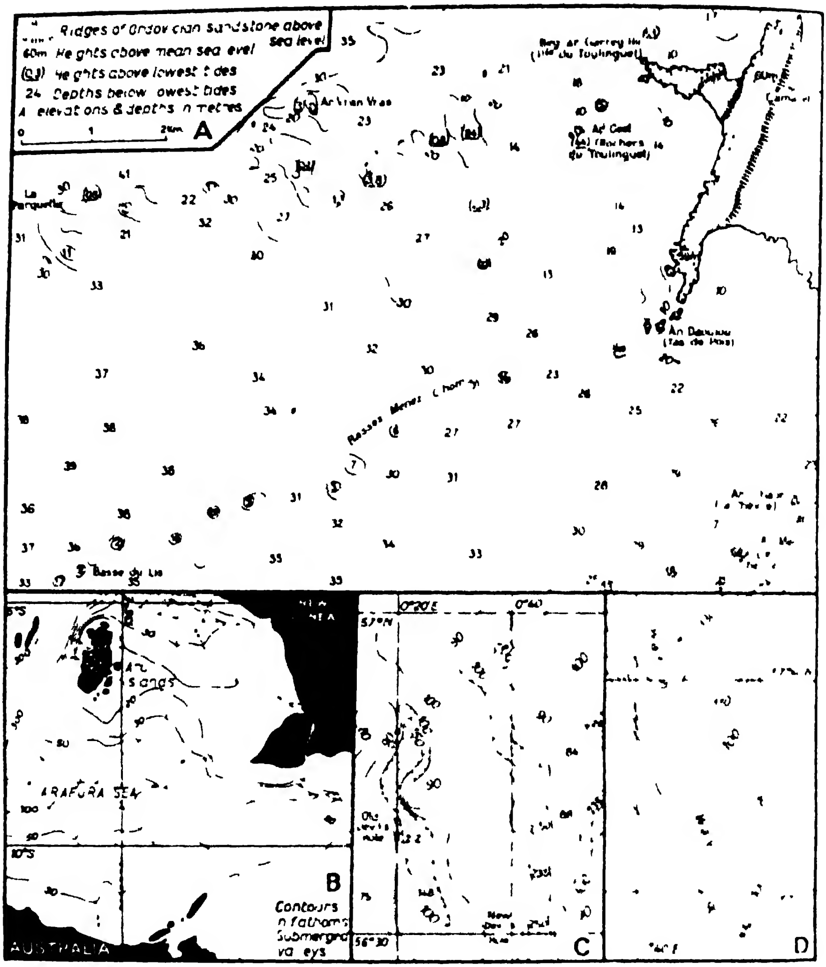

FIG. 30 DROWNED VALLEYS AND STRUCTURAL RELIEF ON THE CONTINENTAL SHELF

A. Off the Crozon peninsula, Finistere (based on French marine charts). B. In the Arafura Sea (after Fairbridge, 1951). C. and D. In the North Sea (after Beaugé, 1937).

has not been confirmed. Submerged plateaux and basins between about 200 and 700 m. occur in certain other regions, e.g. the Bligh and Rosemary banks north-west of the British Isles, north of Newfoundland, and south-west of Formosa. Shepard termed such areas continental borderlands.

These shelves or continental borderlands are generally not plains devoid of relief. Locally they have this character, e.g. off the west coast of India and in the Chukotsk Sea off north-east Siberia, where they are as flat as the plain of the Low Countries. But usually they possess considerable relief, which appears to be a continuation of that of neighbouring land areas.

Exploration by means of self-contained diving suits has revealed vertical cliffs and caves to be common at depths of 0–40 m. off the Provençal coast. Off the west coast of Brittany, the structural trend of the Crozon peninsula may be traced to a depth of about 50 m. (Fig. 30 A). The higher parts emerge in places as rocks, but most of it is always submerged. The ridge of Armorican sandstone at Toulinguet point is continued with the appearance of an *en échelon* pattern in the submerged ridges of La Parquette and La Vandrée for at least 18 km.; the Tas de Pois continues in the Menez C'hom, Lis, and Iroise banks for at least 18 and probably 24 km. There are comparable features off Dinan point and Cap de la Chèvre.

These facts might have been attributed, at least in part, to differential marine erosion working at the present level (Chapter V), if L. Berthois had not shown that wave action is very weak as soon as low tide level is reached. But the submarine valleys on the continental shelf cannot be explained in this way. Such drowned and branching valleys occur in the north Adriatic; they are magnificently developed in the Sunda platform between Java, Sumatra, and Borneo, where there are two former river systems, at depths of 0–100 m., one directed northwards and the other south-eastwards, both continuing the courses of subaerial rivers. Similarly, in the Arafura Sea between Australia and New Guinea, old river systems run westwards and can be followed down to a depth of more than 180 m. (Fig. 30 B). Off the coasts of French Guinea, another pattern of clearly defined channels continues the lines of the coastal rivers and goes down to 70 m. in the channel in front of the Konkouré river. The submarine valley, which leaves the Gulf of Morbihan at a depth of 31 m., reappears to the south-east of Quiberon, where it reaches – 51 m. in the Teignouse channel. Similar valleys exist off the west and south-west coasts of Brittany. The channels of the North

Sea floor provide another example: these channels, the Old Devil's Hole and the New Devil's Hole are long and narrow, and reach depths of 212 and 250 m., although the average depth of the sea-floor is only 85 m. There is also the Silver Pit in the Wash, and to the north, the Swatch Way, a kind of submerged gulf 100–140 m. deep (Fig. 30 C and D), and the Fladen deep (274 m.).

It is almost certain that all these are completely submerged subaerial valleys. In general the pre-Flandrian regression to a depth of 100 m. provides an adequate explanation. Between New Guinea and Australia there may also have been tectonic movement in the Quaternary. In the North Sea the depth of the valleys may be explained, either by a continuation of the subsidence which occurred in the Mesozoic and Tertiary, or, as envisaged by H. Baulig, by isostatic effects associated with the last glaciation. The ice of the last glaciation did not fill the North Sea basin so that the land would possibly have risen along the ice margin. Here the valleys were cut to a depth well below that existing at present. When the ice finally melted, the land sank, i.e. its movements were opposite to those of the parts covered by ice (cf. pp. 46 8). These valleys were formed by the Rhine and its tributaries and later separated into series of depressions by sedimentation, unless they were, according to another of Baulig's suggestions, subglacial valleys carved out during a former and more extensive glaciation.

Off the Norwegian coast (see pp. 159-60) the continental shelf has been heavily glaciated and is crossed by many submerged troughs. Off the coasts of glacial lowlands moraine features are sometimes quite deeply submerged: in the Gulf of Maine many drumlins have been found at a depth of about 180 m. by the surveys of the *Oceanographer*. In the North Sea, three concentric ridges of stony material west and north-west of Jutland have been interpreted by Pratje as the submarine continuations of the Warthe, Frankfurt-on-Oder, and Pomeranian stadial moraines (Fig. 31 C). Off New York the Hudson valley, which was overdeepened to - 90 m. under the city and since filled up, continues across the platform as a submerged valley 50–70 m. deep. It may be of compound origin, in part a glacial valley and in part a proglacial valley.

As a result, the relief of the continental shelf is similar to that of the adjoining parts of the continents, both in its general degree of flatness and in its detail. The relief is largely, but not completely, subaerial in origin; for example, the lines of sand-banks in the Straits of Dover, in the southern North Sea (Fig. 31 B), and in the straits of Banka and

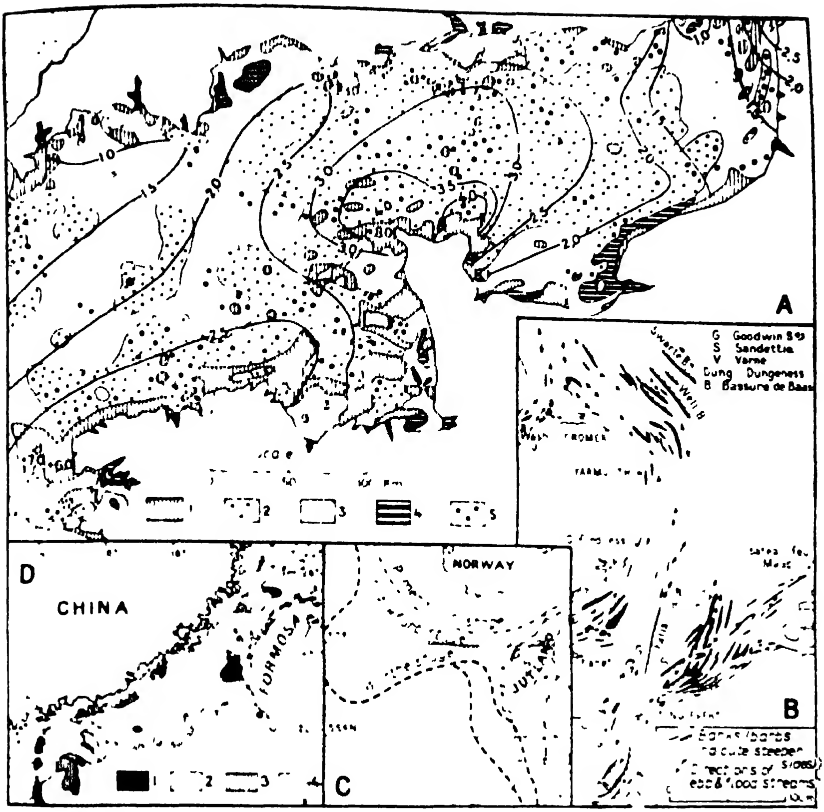

FIG. 31 DEPOSITS IN EPICONTINENTAL SEAS

A. Sediments and tidal current velocities in the English Channel (after Dangeard, Hansen, and Pratje). 1. Rock. 2. Gravel. 3. Sand. 4. Muddy sand or mud. 5. Isolated stones. Lines of equal maximum tidal current velocity in knots are shown, except off Finistère and Cotentin, where the velocities are indicated only by figures.

B. Banks in the North Sea, Straits of Dover, and eastern Channel (after Guilcher). Scale should read km. and not m.

C. Reconstructed positions of stadial moraines based on stony areas in the North Sea (after Pratje).

D. Sediments off the south China coast (after Shepard). 1. Solid rock or stones. 2. Sand. 3. Mud and sand. 4. Mud.

Sunda are forms typical of straits through which strong currents flow. Such features are also found in front of many British harbours (Robinson, 1956). On the Sunda shelf valleys have been scoured out by marine currents in narrow straits. It is quite possible that elongated banks, such as the Parsons, Castor, and La Chapelle banks off western Brittany, which have sometimes been regarded as structural features, are, at least in part, of the same origin as those of the North Sea and the Banka Strait, for the chart shows that in the east-south-east they are formed entirely of gravel, sand, and shells. The fact, noted by Bourcart and Marie, that trawlers have dredged up compound branched polyps is not complete proof that the banks are rock.

Thus, the continental shelf is a region where continental and marine influences have acted alternately during successive emergences and submergences. But the realization of this gets us little nearer to understanding the origin of the shelf itself.

B. RELIEF OF THE CONTINENTAL SLOPE: SUBMARINE CANYONS

The continental slope, where it is relatively well known, is both steep and, in detail, sinuous in plan, although its general form may be simple. It is characterized by great gorges with steep and sometimes vertical slopes, some of which, e.g. the Cape Breton trench, cut deeply into the continental shelf itself and end in marked amphitheatres. These trenches have been called submarine canyons (Fig. 32). Usually their long profiles show breaks of slope but no reversed slopes. This profile is usually much steeper than that of subaerial valleys, except in certain high mountain regions. According to Shepard and Beard the average slopes of 102 canyons known in 1938 were as follows: 11·62 per cent in the upper, 6·63 per cent. in the middle, and 4·76 per cent. in the lower sections.

The majority of canyons do not appear to be the continuations of subaerial valleys, but there are exceptions, for canyons are found off the mouths of the Indus, Ganges, Hudson, and especially the Congo, into the estuary of which the canyon penetrates deeply, a very rare occurrence. The Tagus canyon lies off a probable former mouth of this river south of the modern estuary. The Cape Breton trench lies off one of the old mouths of the Adour. But the profiles of the canyons are not continuous with the profiles of the subaerial valleys. In California, the Monterey canyon continues the Salinas valley but the canyon has a

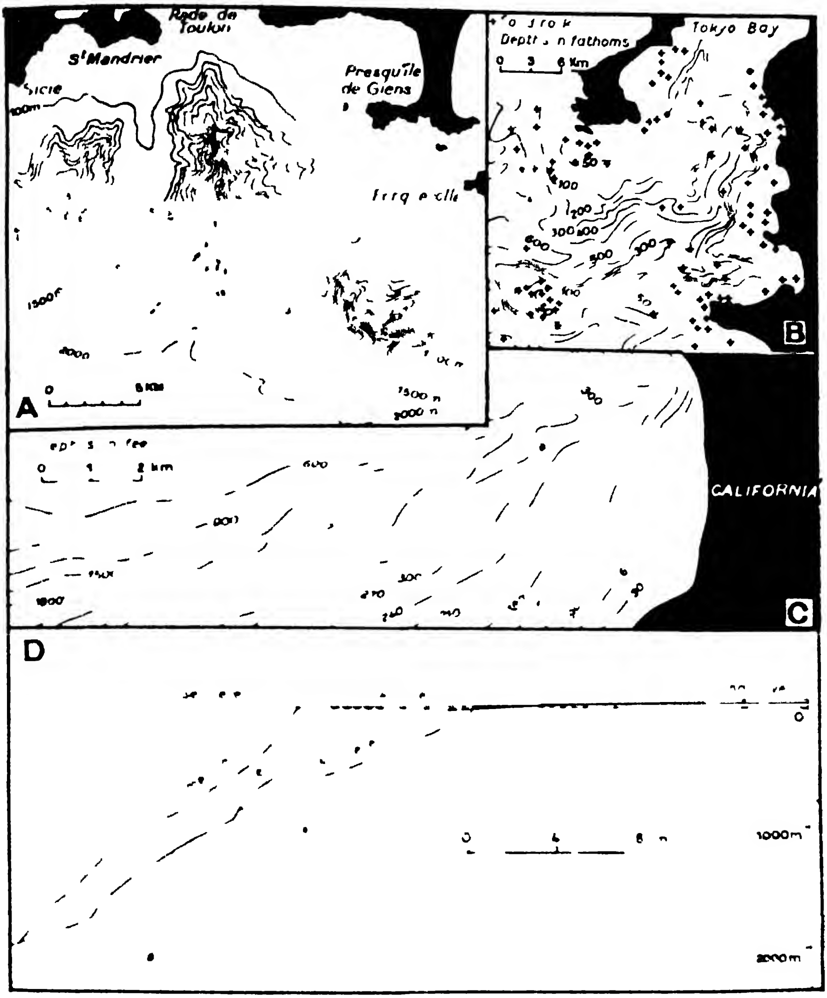

FIG. 32 SUBMARINE CANYONS

A Sicié, Toulon and Porquerolles canyons Provence (solid lines after Bourcart, Houot, and Lalou pecked lines after Marti and Anthoine B Canyon in Tokyo Bay after Shepard C Redondo Canyon California (after Shepard and Emery D Comparison of long profiles of Californian canyons and subaerial valleys leading into them (after Crowell).

steep, constant slope, which is not a logarithmic curve, the break of slope near the coast being very abrupt (Fig. 32 D). The canyons, which are apparent extensions of rivers, are longer than the others (134 km. in the case of the Congo canyon, and 94 km. for the Indus canyon) and their longitudinal profiles with an average slope of 1·7 per cent. are less steep. Canyons lying off islands have the steepest profiles of all with an average of 13·8 per cent.

Cross-section of canyons are usually V-shaped like those of the Cévennes valleys. In detail there are breaks of slope which are often of considerable magnitude. Shepard has superposed the cross profiles of the Monterey canyon, California, and the Colorado canyon, and shown that they are very much alike. The depth of the canyons is usually several hundred and sometimes more than a thousand metres, a fact which clearly distinguishes them from the other submarine valleys of the continental shelf. Comparable subaerial features are by no means unknown for the Colorado canyon is 1,830 and the Snake River canyon 2,370 m. deep. The Mississippi canyon in the form of a flat-bottomed trough is exceptional.

The plan is often dendritic, like that of a subaerial valley system, e.g. off the coast of Provence (Fig. 32 A). However, Kuenen, Woodford, and Crowell have emphasized certain differences between canyons and subaerial valleys. As yet no notably winding canyons or meanders of small radius have been found: the canyons are not really straight but they are straighter than the average subaerial valley, while irregularities in the long profile are greater in the submarine canyons.

The distribution of submarine canyons is almost world wide. They are particularly numerous off the east coast of the United States between Canada and Norfolk, Virginia, off the Californian coast, the Pacific coast of Mexico, and the Mediterranean coast from Cape Creus to the Italian Riviera, where they are in the outer edge of the continental shelf in the Gulf of Lion, but quite near the coast from Toulon eastwards. But they occur commonly in other regions: in the western approaches to the Channel, off Algeria, Senegal, Chile, New Guinea, the Philippines, Japan, the Aleutian Islands, Ceylon, and Zanzibar. Very recently, Carsola has mentioned canyons cut into the continental slope of the Arctic in the Beaufort Sea. They seem to be absent from the west coast of the Gulf of Mexico and between Florida and Norfolk. Hundreds of canyons are now known.

They generally reach a depth of 2,000–3,000 m. and those which have been sounded in their lower part end in areas of hummocky relief.

C. NERITIC AND BATHYAL SEDIMENTS

The continental shelf is an area of shallow water or neritic sedimentation, while the continental slope is an area of deeper water or bathyal sedimentation. The latter extends, according to most current definitions, down to 1,000 m. but, as Kuenen observes (*Marine Geology*, p. 322), it may be advantageous to continue it to 2,000 m.

If sea-level remained constant for a long while, and if waves were the only important factor in sedimentation, the calibre of the sediments should decrease away from the land. Sediments derived from the land should be arranged in such a fashion in places where the slope decreases away from the land, because of the reduction of movement with depth. This would result in what Baulig calls (*in litt.*) the profile of equilibrium of the continental shelf. Some authorities have gone further and stated that such was the arrangement of sediments on the continental shelf. Conclusions have often been drawn from the nature of the sediments about the depth of water in which they were deposited.

In reality the distribution of sediments on the continental shelf is very complex. As Twenhofel (1939) says, 'The coarsest sediments tend to be nearest the shore, but there are many exceptions, and on many shores fine sediments are nearest the shore and the coarsest sediments are seawards' (*Principles of Sedimentation* p. 119). Pebbles have been found at great depths in a variety of places. Off southern California, they have been dredged up from all depths between 18 and 900 m. Similarly, off western Europe the *President-Théodore-Tissier*, the *Quentin-Roosevelt*, and the cable ship *Emile Baudot* have found pebbles at considerable depths: at 80 m. off the Lizard; 105 and 110 m off the Scilly Isles; while very large pebbles were found at a depth of 160–175 m., a little farther to the south-west. At a depth of 140 m. south-west of the Chaussée de Sein a pebble of sandy shale was found. These pebbles are well rounded and very varied lithologically. They cannot have been formed at the depth at which they are found. Submarine currents may form ripple-marks on sandy bottoms: they have been photographed on George Bank, New England, at a depth of 225 m., and many are known at greater depths in certain straits (see p. 81). But to move sand and to erode pebbles are two quite different things. Moreover, these pebbles, like those which are exposed among generally angular cobbles at very low water on the coasts of Brittany, are covered with calcareous algae which cannot withstand erosion. At most it is probable (Berthois) that they are occasionally rolled over since they

have calcareous algae on all sides. Sometimes they are so abundant that they appear to form submerged shingle ridges.

The distribution of sediments in the English Channel is very irregular (Fig. 31 A), as was shown by Dangeard and later by Berthois. In the eastern part of the English Channel numerous angular flints occur at a depth of 50 m. in 59° 24′ North and 2° 21′ West the following rocks have been found: gneiss, felspathic sandstone, iron-cemented sandstone, glauconitic sandstone with a calcareous cement, foraminiferal limestone, oolitic limestone, limonite nodules, and fossil wood probably of Wealden age (Berthois). In the Rade de Brest each little basin has its own particular grade of material, and the particles are almost unworn (Berthois).

An even more interesting fact is the distribution of sand and mud on the continental shelf. The existence in the Bay of Biscay of a large mud-bank, forming a broad strip between the Gironde and Penmarch, and flanked by sandy areas, has been known for a long time. The sand appears to extend continuously beneath the mud-bank, which represents very localized deposition. It should be added that there are areas of mud nearer the coast, notably in the Mor Bras between Houat, Hoedic, and the coast of Morbihan, so that the succession outwards from the coast is mud, sand, mud, sand.

Such anomalies are very numerous, as Shepard has shown. He has drawn attention particularly to the south China coast, where a zone of sand appears outside a broad strip of mud flanking this ria coast (Fig. 31 D). Similarly, on the north-east coast of South America from the Orinoco to the Amazon, vast mud-banks border the coast for more than 2,000 km., and continue into the deeper parts near the coast where they are red-brown in colour and have a high content of lime (6–60 per cent.) and organic matter (30 per cent.). On the seaward parts of the continental shelf and on the continental slope the same mud occurs, but is interrupted by areas of sand or muddy sand. South of Cape Cod, compact shelly material covers the bottom between 35 and 45 miles from the coast at depths of about 90 m., but soft mud appears at depths of about 100 m. between 45 and 70 miles from the shore, between 75 and 85 miles from it, in depths of about 110–120 m. the grain size of the sediment and its compactness again increases in a zone where no deposition seems to be taking place at present (Northrop).

Organic sediments are also important on the continental shelf. Far from being limited to the deepest parts of the seas, these sediments are commonly found in Brittany just below the level of the lowest tides and

are derived from the calcareous algae which cover all the rocks, and from the shells of mussels, and sea-urchins. In places they completely cover the sea-floor. Breton farmers know the deposit well, for they have used this '*maërl*' or '*skotailh*', which can be seen heaped up on the quays at places such as Quimper, for centuries for fertilizing fields deficient in lime. In the Mediterranean such deposits are equally important; dredging by the *Elie-Monnier* in August 1950 off Saint-Tropez and around the Ile du Levant has shown that the bottom between 100 and 150 m. is in places entirely covered with Bryozoa. These deposits are the '*madrépores*' of the marine charts. In coral seas both sands and muds are exclusively organic.

Solid rock also outcrops over extensive parts of the continental shelf. Dangeard has greatly increased our knowledge of this. The outcrops of solid rock are so numerous in the English Channel that King has been able to make a geological map of the bottom of the Channel. Dangeard states that rock may occur well out to sea and at great depths. This is not peculiar to the English Channel, for such rocky bottoms, sometimes called hard grounds or hard bottoms, have been noted in different regions by Twenhofel (*Principles* . . . , first edition, 1939, pp. 253–4), notably in Chesapeake Bay, the Straits of Gibraltar, and around the British Isles. Off California, solid rock alternates with sand on the outer part of the continental shelf. Off the South China coast rock crops out in many places in both sand and mud zones.

These facts are not easy to explain. If we consider only the Flandrian transgression and wave action to be the operative factors, we should have either a continuous basal conglomerate covered by finer sediment in its lower parts and in coastal marshes or the rising sea should have pushed its shingle beach up as it advanced and afterwards deposited other sediments below it in deeper water. Neither of these simple distributions is usually found, except occasionally in vertical sections through marshes.

Submerged pebbles may be due to several causes. Isolated pebbles may have been dropped by floating ice in the Quaternary, according to the theory, popular in England, for explaining erratic pebbles in raised beaches (p. 26). This theory may apply in certain cases: at present material is being deposited by Arctic ice between Greenland and Spitzbergen. But it does not account for the presence of submerged ridges, which must have been formed and then abandoned by the rising sea. Why this should be so is not clear. A better knowledge of the morphology of the sea-floor may reveal the influence of rock ridges.

The causes of sand zones seawards of mud zones, other than modern coastal mud-banks, like those of the Bay of Biscay and south of Cape Cod, are still very obscure, and perhaps related to hydrodynamic conditions. Writers who describe these sand zones think that they are generally old, so that the conditions of deposition are very difficult to determine. It may be easier to discover why the mud, which is usually considered to be of recent origin, does not cover them completely. Is the great mud-bank of the Bay of Biscay, as H. Baulig suggests, being actively built up seawards? Or is it stable in position and to be explained by hydrodynamic conditions? It seems likely that many irregularities of deposition must be due to submarine currents. Although currents, other than wave currents, are not important in coastal geomorphology, they may be relatively more effective on the continental shelf. The sand ripples which have been photographed in deep water prove the importance of currents, but it is difficult to establish their significance in every case. Pratje recently used comparable ideas in trying to explain the transition from sands to gravels in the Channel by an increase of the tidal currents (Fig. 31 A). There is no doubt that rocky bottoms usually owe their existence to the fact that they are swept by currents, for example in the Channel and the Straits of Gibraltar. Others may result from the fact that the deposition of sand and mud has been insufficient in thickness to cover the pre-existing relief, a probable example being those of Iroise in west Brittany.

In certain cases the origin of the sediments has been established. In the southern part of the North Sea it has been shown by the heavy minerals and by the shape of the sand grains, that practically none of the material is of distant origin, because the marine sediments have the same characteristics as those of the adjacent lands. Some of the sediments on the Flanders banks may have been brought into the Channel from the North Sea, when the Channel was dry land in the penultimate glaciation, by the overflow stream responsible for breaching the Straits of Dover. Later the sea has tended to carry them back to the area from which they came, without mixing them with those lying off East Anglia. Further studies of this type are much to be desired.

Sedimentation on the continental slope, in the canyons, and at their mouths is as interesting as that on the continental shelf. On the continental slope rock outcrops frequently. It usually alternates with fine muddy sediments, for example, south of Cape Cod, off south California, off the mouth of the Columbia river, and off the Atlantic coast of France. But these unconsolidated sediments are not very thick. The

nature of the material is very variable, as, for example, south of Cape Cod. A fairly common deposit near the edge of the slope, but occurring at any depth between 100 and 2,500 m., is a green mud, the colour being due to the presence of glauconite. Lime forms about 50 per cent. of it and it also contains nodules of calcium phosphate.

Rock often outcrops on the sides of the canyons (Fig. 32 B) although the bottom is usually covered with sediment Muds are often found in them but sand, which is angular in the Hudson canyon, and even pebbles may occur. These materials have often been dredged from the Scripps and La Jolla canyons in California and also from the canyon of St Tropez in Provence. In the latter quartz pebbles, some of them very well rounded and often encrusted with and embedded in sandstone and mixed with gravel, were found between 200 and 300 m. (pl. VI D).

At the heads of the canyons near La Jolla, where the slopes are covered with sand, Shepard has shown by careful soundings that changes of as much as 6 m. in three years may occur. These are explained by alternations of deposition and landslides. Finally, sediments, abnormal in composition and arrangement and associated with irregular relief, i.e. alternations of beds of sand originating from the canyon and of deep-sea clay, occur in the North Atlantic, especially at the mouth of the Hudson canyon. They may be caused, according to Ericson, Ewing, and Heezen, by the sliding of deposits from the floor of the canyon to greater depths. In California there is an extensive delta at the mouth of the Coronado canyon off San Diego.

D. NATURE AND ORIGIN OF THE CONTINENTAL SHELF AND THE CONTINENTAL SLOPE

What is the nature of the continental shelf beneath its covering sediment? The classical theory, which was based, for want of adequate means of investigation, on deduction alone and not on observation, regarded the outer part as being composed of consolidated sediments, comparable with but older than those on the surface of the shelf, and the inner part of rocks like those of the adjacent lands thinly covered with sediments. The inner part was, thus, an abrasion platform, and the outer part a constructional form built up of material eroded during the formation of the inner part and material brought by rivers. The continental slope is thus the outer slope of this embankment of talus (Fig. 3 A). On prograding shores and where large rivers debouch, e.g. the north coast of the Gulf of Mexico, there would be no abrasion platform, but only the outer, constructional part of the shelf.

This idea is not entirely wrong but needs modification, and even then is not valid for all continental shelves. It involves several serious difficulties.

In the first place, it is highly unlikely that, apart from a few localities, the inner part of the shelf has been formed by wave abrasion if it is more than several kilometres wide. This point is discussed in the first part of this book. In general, marine abrasion is of limited extent.

The uneven relief of the platform does not favour the idea that it has been simply formed as a delta. Low sea-levels in the Pleistocene may have resulted both in some dissection of the platform and in irregular glacial deposition. A further objection to the delta hypothesis is found in the outcrops of crystalline and Palaeozoic rocks found by Dangeard in the Channel, and of granite in certain Californian canyons.

Dredging in canyons and sounding by geophysical methods (p. 197) allow us to get nearer to the facts. Both show that the origin of the shelf is not everywhere the same: in some regions it is a constructional form, but the sediments have been downwarped through subsidence, probably caused by their weight.

The best example of a constructional form, with subsidence of the sediments, is the shelf off the east coast of the United States. Numerous seismic soundings have shown that the Appalachian rocks, which disappear under the Mesozoic and Tertiary rocks at the Fall Line, are affected by a flexure which lowers them to about 900 m. below sea-level at the mouth of Chesapeake Bay, and to 3,900 m. near the outer edge of the continental shelf. Moreover, it has recently been established by seismic work off New York that the Appalachian rocks rise again seawards, a fact not revealed by earlier soundings (Fig. 33 B and C). The same state of affairs has been found south of Nova Scotia and south of the Grand Banks of Newfoundland by Ewing's team. These facts are confirmed by dredging which has established the presence, as far as the continental slope, of a mass of rocks ranging from Cretaceous to Quaternary in age, and especially by the boring at Cape Hatteras which reached the Appalachian rocks. The latter proves a progressive thickening of the Mesozoic, Tertiary, and Quaternary strata, as these are all thicker than in two other borings made farther to the west (Fig. 33 A). The constructional nature of this shelf and its downwarping by a sort of geosynclinial subsidence are, therefore, definite facts.

The well-known coastal shelf in the north of the Gulf of Mexico is of the same type, and has been formed of a great mass of material brought down by the Mississippi, which has been at least partly responsible for

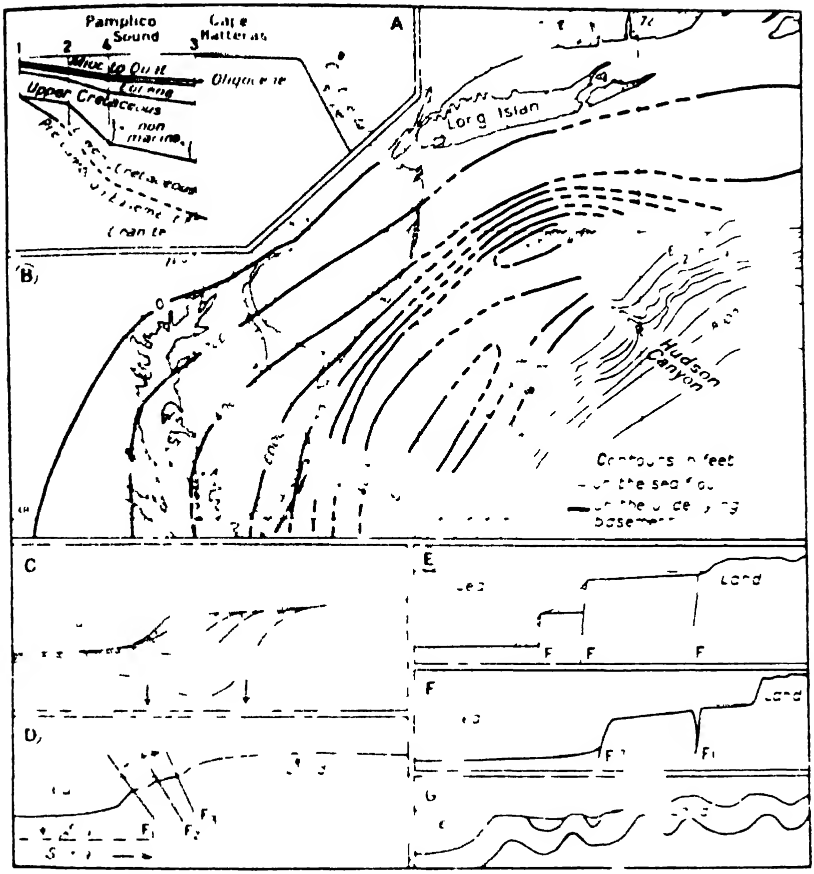

FIG. 33. STRUCTURE OF THE CONTINENTAL MARGIN

A. Cape Hatteras borehole (after Swain and Kuenen, 1950). B. Structure of the continental border off New York (after seismic work by Emery, Worzel, Steenland, and Press, 1950). C–G. Possible types of continental border. C. Subsiding type as in north-east America. D. Flexured type as in Africa. E. Faulted type as in Queensland. F. Fissured type as in Norway. G. Folded and possibly flexured type as in Algeria.

its subsidence (p. 117). Borings in Louisiana show the same thickening of the sediments near the sea as at Cape Hatteras. It is more than likely that the region near the mouth of the Congo is also of this type the Cretaceous and Tertiary rocks of the coastal plain are some 2,500–3,000 m thick and rest on a surface sloping towards the sea The evolution here has perhaps been more complex, since Veatch describes the Cretaceous and Tertiary rocks as folded, faulted, and eroded A downwarping of the Palaeozoic basement is also likely in the western approaches of the English Channel, which seem to be a post-Carboniferous geosyncline in which the strata increase in thickness towards the edge of the continental slope (Day, Hill, Laughton, and Swallow, 1956) Limestone, probably of Oligocene age, has been found on the same continental slope at a depth of 2,500 m Bourcart and Marie, 1951

In certain regions it is possible that there has been a flexure but no thick accumulation of sediments Bourcart and Jessen suggested the idea also adopted by Umbgrove, that the continental border is a zone of flexure (p 49) These authors find flexures in many regions, even where there are no signs of any loading of sediments on the shelf Gravity measurement made by Vening Meinesz in a submarine show positive anomalies off many coasts According to the geophysicists the earth consists of an outermost layer of sial sedimentary and granite rocks) and a layer of higher density, the sima (basic igneous rock below The positive anomalies found by Vening Meinesz must, therefore, indicate a great thickness of sima on the sea-floor, while the neighbouring land is formed of relatively light sial resting on the sima

The unequal distribution of sima and sial along the coasts might cause deep convection currents, leading to periodical movements of the flexure at the edge of the continents All this implies a doming of the continent and the transference of sima beneath the continent compensated to some extent by surface erosion Other hypotheses to account for the flexure have been suggested Jessen assumes a periodical cooling of the earth, more marked under the oceans than under the continents This causes subsidence under the oceans and movement of material to beneath the edge of the continents, which are thus warped up A marginal flexure appears probable on certain African coasts, notably off Angola This would agree with the ideas of Dresch, who thinks that the edges of Africa have been recently warped up and that the rivers cutting through them are antecedent But a simple continental flexure only explains the nature of the ocean continent boundary and not the boundary between continental shelf and slope We should have to

postulate, as Bourcart and Jessen do, at least two phases of flexuring, either separated by a long period of subaerial erosion responsible for the continental shelf, or characterized by a migration of the line of flexure towards the land. More phases of flexuring could lead to the formation of the more deeply submerged continental borderlands (Fig. 33 D).

A third possible type of border may be caused by step-faulting (Fig. 33 E) We have seen (p 164 that Steers considers that the platform on which the Great Barrier Reefs are built is probably of this type. The coast of the Red Sea is partly and possibly wholly of this type. As on the flexured margin, there may well be a compensatory flow of sima from the ocean floor to beneath the continent If the fault nearest the continent has a throw greater than 300 m the downthrow side will be so deeply submerged that a faulted coast without a continental shelf will be formed As in the case of a flexured margin, sedimentation plays no great part in the formation of the continental shelf, but may have the effect of smoothing out the irregular relief of the continental shelf The shelves bordering mountainous regions, discussed by P Birot (p. 205), may well be of this type, especially that off Galicia

The fissured border, to which O and H Holtedahl have drawn attention Fig 33 F, is really a variant of the step-faulted borders Along the old shields of Norway, Labrador, and Greenland, which have undergone recent uplift, O Holtedahl has shown that outside the zone of flat-bottomed depressions which cross the shelf, there is often a furrow parallel to the coast In Norway, this furrow in places is more than 200 m deep, and separates the strandflat from the outer part of the continental shelf which may be at a depth of 50–200 m It must have been modified by fluvial and glacial erosion, but as it is perpendicular to the direction of movement of the ice and of no great width, it has probably been modified less than the transverse troughs It is probably the result of fissuring associated with the uplift of the land It seems likely that this is genuine tectonic relief, compared with the fjord coasts which are more like fault-line scarps On the other hand, the continental margin off southern California probably consists of horsts and grabens parallel to the coast, but, here, no glacial action occurred

A folded border may occur where the folds are parallel to the coast, and where the folds are lower on the continental shelf than on the continent (Fig 33 G) The continental shelf off eastern Algeria and northern Tunisia is probably of this type, according to Bourcart and Glangeaud (1954), for it consists of a series of parallel crests running south-west–north-east The oblong basins between them seem to be

partly filled up with sediments. These folds are possibly combined with a marginal flexure.

In all these cases it is not necessary to presume that the flexure is recent unless the submarine canyons lead one to that view (see below). The flexure may be old, or possibly have undergone movement over a long period.

It would be wise to come to the same cautious conclusion as Kuenen (*Marine Geology*, p. 169) that the problem of the shelf is still far from being definitely solved. The problem may be solved to some extent either now or in the future by the study of submarine canyons, which remain to be examined.

E. ORIGIN AND EVOLUTION OF SUBMARINE CANYONS

Canyons cannot be older than the continental shelf and slope, since they are cut into it; the stratigraphy of the outer part of the shelf may be discovered from their steep sides. The age of the rocks dredged from these walls varies from place to place. In the Banyuls canyon Bourcart found fossiliferous Middle Miocene rocks and probably older unfossiliferous rocks; in the Sicié and Cassidagne canyons, off Cassis in Provence, fragments of phyllites comparable with those of the ancient massif of Sicié were found. It is, therefore, theoretically possible that the latter two canyons are much older than the Banyuls canyon; but the discovery of pebbles in the Saint-Tropez (p. 217) and certain other canyons suggests that all the canyons of the French Mediterranean coast are very recent in origin. J. Bourcart dates the sandstone associated with the pebbles as Lower Quaternary. The Cape Breton trench is cut, at least in part, in Eocene rocks. The canyons of New England are cut into the Pliocene, and those of California into late Tertiary beds. It seems, therefore, that we must search for an explanation or explanations, which attribute no great age to these canyons.

In spite of the amount of writing on the subject, the origin of canyons remains shrouded in mystery. The theory of a glacial origin is, however, untenable because of the distribution of the canyons, their V-shaped cross-section and their depth. They are not due to faulting or to rifting except possibly the unusual Mississippi canyon which has the appearance of a rift-valley. As with fjords, it is possible that faults have guided their erosion, but the form of the canyons, which are obviously erosional features with generally concave profiles, few reversed slopes, and V-shaped cross-sections, cannot be explained structurally.

The theory that they were formed by submarine springs has been put forward in two forms: mechanical sapping by W. S. T. Smith and hollowing out by solution by D. W. Johnson. These theories are perhaps possible for some canyons. The efficacy of such processes has been recognized on the continents, where they cut short canyon-like valleys. Solution would be more effective if the springs were hot, and could result in a relatively rapid hollowing out of non-calcareous rocks. The theory of hot springs was put forward to explain the Cape Breton trench, but the *Président-Théodore-Tissier* eventually found that the temperature of the water there was normal. It must be said that the number of canyons possibly owing their origin to solution by submarine springs is very small. For such action, there must be a series of conformable beds dipping seawards, so that canyons off old massifs and in folded regions, e.g. Provence, California, and Japan, cannot be formed in this way. Further, springs would require a very long period to cut long canyons so that the hypothesis cannot apply to those off the east coast of the United States which are cut into Pliocene beds. Mechanical erosion by springs would be more likely to form isolated closed depressions than canyons.

The simplest explanation is that of subaerial river erosion. But this implies either an enormous eustatic lowering of sea-level, or vast tectonic movements of the continental margin. The idea of a great fall in sea-level was held for a time by Shepard, who postulated a glacio-eustatic lowering of 1,200 m. and complementary isostatic adjustment of 600–800 m. There are many objections to this theory, especially the enormous biological changes which should have resulted from such a shrinking of the oceans and the accompanying great increase in salinity; and the impossibility of explaining in this way the Mediterranean canyons as the Mediterranean could not have been lowered below the level of the sea-floor in the Strait of Gibraltar. Such a fall in sea-level would have had glacial implications and would have created land bridges, for which there is not a shred of faunistic evidence. Shepard abandoned this explanation for another in 1951 (see below).

A diastrophic hypothesis is adopted by Bourcart and agrees with Jessen's theory. According to Bourcart, canyons are Quaternary river valleys submerged by a recent movement of the continental marginal flexure, as shown in Fig. 34. This hypothesis, which is in accordance in Provence with the borings recently carried out into the Rhône delta (Beaufort and others, 1954), has the advantage of offering an explanation of the longitudinal profiles of the canyons, which are abnormally

steep for river valleys: the profiles would not be original but steepened during the flexuring. Multiple movements of the flexure could account for the stepped profiles of the lower parts of some canyons such as that off the Congo: thus the number of canyon steps would represent the number of movements of the flexure. The canyon would, in fact, be polycyclic. The explanation would be easier to accept if canyons of apparently recent origin were all concentrated in regions where large-scale earth movements are known to have occurred in the late Tertiary

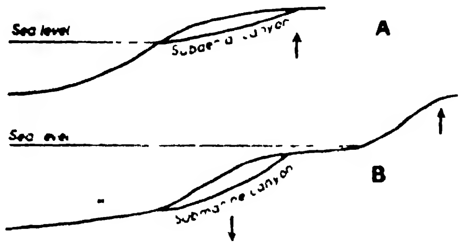

FIG. 34 STEEPENING OF THE LONG PROFILE OF A CANYON BY PRESUMED FLEXURING

and Quaternary periods, as in the Mediterranean and Pacific regions, where canyons are found off the Philippines, Japan, the Aleutian Islands, California, western Mexico, Chile, and north-east New Guinea, or if the movement along the flexure had continued for a long time, as Jessen thinks happened in the Congo. It is not possible to invoke either of these theories in some regions, notably on the Atlantic coast of the United States, where canyons are cut into the Pliocene and where there is no evidence of recent movements of any importance other than isostatic subsidence under the weight of sediments. Glacial isostatic movements are quite insufficient as an explanation.

The fact that most canyons are not connected with continental river systems or with shallow submarine valleys on the continental shelf is another disturbing fact. The heads of the canyons are found at very variable depths even within one region such as California. Rivers are known, which begin in steep amphitheatres cut into a plateau and which have no connexion with the drainage of that plateau: examples occur in the Cévennes where the ravines flowing to the Mediterranean have no connexion with the headstreams of the Garonne. But in such cases the river system of the plateau flows in the opposite direction from that of the ravines. On the continental shelf, there ought to be, therefore, a non-entrenched drainage directed towards the continent, but this is

not the case. There are, furthermore, no signs that such a system once existed but has since been either disrupted or reversed. On the contrary, there is every reason for believing that coastal river systems, with a few exceptions, had the same direction in the Quaternary as they have today, even in places where they do not end in canyons.

Further, even when there is a subaerial valley at the head of a canyon, the subaerial theory encounters difficulties. Thus, in the case of the Salinas valley and the Monterey canyon (Fig. 32 D) the abrupt break of slope, which separates the two, occurs in soft sediment. Woodford observes that even if rejuvenation could produce an initial break of this kind in soft rock, the break of slope would change rapidly into a broad convexity. It is, therefore, unlikely that the Monterey canyon has been formed by the rejuvenation of the lower part of the Salinas valley.

Another explanation invokes turbidity currents and submarine landslides. Daly and Kuenen regard turbidity currents as having formed the canyons. Shepard has recourse to them in order to explain the fresh appearance of canyons which were originally formed by another process.

The turbidity current theory may be summarized as follows. During the Quaternary glaciations, sea-level was lowered eustatically almost to the average height of the heads of the canyons, storms drove the surface water towards the land and stirred up great quantities of fine sediment on the beaches exposed by the fall in sea-level. Water charged with this mud in suspension formed compensatory currents which flowed down the continental slope and hollowed out the canyons.

This theory, put forward by Daly and Kuenen, rests upon certain observed facts and experiments, which in themselves are incontestible. The density of the water of turbid rivers makes them plunge downwards where they enter lakes. They follow ravines in the lake floors, which, however, may be due more to an absence of sedimentation than to real erosion. The experiments made and filmed by Kuenen at the Geological Institute of Groningen show clearly that turbidity currents may attain quite high velocities. Landslides on the continental slope may be transformed by the addition of water into turbidity currents. If their relative density is 2 and their thickness 4 m., they reach a speed of 3 m. sec. in a 3° slope. The velocity is greater in the lower layers of the current than at its surface where there is turbulent mixing with the clear water above. Thus erosion of the bottom is facilitated. The properties of these currents seem to be the same in salt water and in freshwater. Once the canyon is formed, frequent landslides are

presumed to occur on its walls, and most of the material would be set in motion by such slides. Turbidity currents would only cause a certain amount of vertical erosion and thus allow landslides to continue on the walls of the canyons. They would also ensure the removal of material from the canyons.

It is quite likely that such currents and landslides still occur today in certain canyons. Shepard's measurements at the heads of the Californian canyons have shown that landslides occur (p. 217) and the areas of irregular relief at the mouths of the American Atlantic canyons may probably be interpreted as waste fans formed by the mud-flows. If all the canyons were cut in unconsolidated deposits, this would be a satisfactory explanation. We may assume that the slides or flows in certain cases are initiated either by waves caused by earthquakes or by submarine earthquakes which may cause considerable changes in submarine relief as has been shown in Japan. This is similar to Bucher's view, although some of his premises are difficult to admit.

Unfortunately, canyons are often cut in hard compact rocks, as in Provence and Corsica and in places off California. These present difficulties, to which it is impossible to give a satisfactory solution.

Landslides and turbidity currents satisfactorily explain the fact that the canyons are not filled up. Kuenen's successful experiments have taught us a great deal about the detailed mechanism of these processes. They help to explain numerous aspects of present and past submarine sedimentation, e.g. underwater slides on the flanks of volcanoes, varved clays, slumped structures and graded bedding in the rocks of such places as New Zealand and Wales (see also Chapter VIII, pp. 252–3). This is why Shepard admits their great efficacy, although he refers the cutting of the canyons to various earlier periods and considers them to be river valleys submerged by subsequent earth movements. This leads us back to the problem of diastrophism and its attendant difficulties.

In one of his recent publications, Kuenen (1952) puts forward the idea that there are two types of canyons at least. Canyons like those off the west coast of Corsica and off the Provence and Nice coasts, which are cut into hard rocks, are submerged subaerial valleys. Others like those of New England may be formed by landslides and turbidity currents or be old canyons from which unconsolidated material has been excavated by the same processes. This seems to be the most reasonable conclusion that can be reached at present.

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Chapter VIII

THE DEEP-SEA FLOOR

A. RELIEF AND STRUCTURE OF THE OCEAN DEPTHS, RIDGES, SEAMOUNTS, AND TABLEMOUNTS

Classification of forms

Although those parts of the oceans between 2,000 and 6,000 m. comprise four-fifths of the total area occupied by oceans and seas, they are far less well known than the continental shelf and slope, so that we are justified in considering them somewhat more briefly. Extensive plateaux, such as those of Fiji, the Bahamas, and Socotra, rise in places from the depths of the oceans. We will not deal here with the parts above − 2,000 m. except where they are small isolated areas, for these often present the same problems as the adjacent deeps.

Within this zone, we may distinguish three types of relief.

(*a*) *Basins and troughs.* Basins often have little relief, e.g. the middle of the north Pacific basin where a number of lines of soundings are available. A comparable basin is that of the Arabian Sea, the bottom of which is almost level at a depth of 3,400 m. Other basins have more diversified relief, e.g. the north-west part of the North Atlantic basin, where a number of seamounts of varied form and height occur. In 39° N. and between 61 and 65 W. a group of such mounts rise up to − 1,440 m. There are certain resemblances between the basins of the Atlantic and Indian Oceans: both are split in two by a central north south ridge. In the Atlantic the two series of basins are quite separate from the Antarctic as far north as Jan Mayen Island, except near the Equator where they are joined by the Romanche trench. Farther north, conditions are different and the basins of Greenland, Spitzbergen, and North Spitzbergen stretch right across the ocean. North of them lies the Arctic basin, in which very important researches carried out by Russian scientists have recently found two basins separated by a high ridge connecting Ellesmere Land with the New Siberian Islands; a third smaller basin lies to the north of Alaska. In the Indian Ocean the

two series of basins are discernible as far north as the latitude of the southern tip of India: farther north the western basin is very narrow. The basins of the Pacific are not arranged in the same way. Three large basins can be recognized, in the north, north-west, and south-east. The last, which lies close to South America, is less deep than the others and resembles a ridge in some ways. The part of the Pacific west of a line joining New Guinea and Japan is much more broken up and ought not to be regarded as a part of the true Pacific but as a zone of bordering seas. Three basins may be distinguished in the Mediterranean and four or five in the Caribbean Sea.

(*b*) *Ridges and rises.* Such features border the basins and are usually less than 2,000 m. deep. Their relief is marked and they generally follow a constant direction for long distances.

The main ridges are those which divide the Atlantic and Indian Oceans longitudinally, and the Lomonossov ridge in the Arctic. The mid-Atlantic ridge, which is S-shaped in plan like the Atlantic itself, starts between Bear Island and Jan Mayen in about 73° N. 5° E., according to recent German work (Fig. 35). All the islands of the central Atlantic are situated on it. It trends north-east–south-west as far as Jan Mayen, then north–south to a position east of Iceland. It is continuous from there to 55° S., except for the break caused by the Romanche trench. Between 30 and 40° N. it is well known from the work of the American ship *Atlantis* (Fig. 36 B). The central part is formed of troughs and crests parallel to the general direction of the ridge. In places the crests reach to within 1,440 m. of the ocean surface and the troughs are sometimes 1,500 m. lower. Flanking this zone, a series of plateaux occurs between 2,880 and 4,500 m. Between these plateaux and the Nares trough, which extends from here to the Bermudas, is a mountainous zone. The German ship, *Meteor*, discovered a similar series of crests and troughs in the southern hemisphere.

The central ridge of the Indian Ocean is Y-shaped in plan. The eastern arm of the Y starts near Karachi and consists of two ridges separated by a deep trough: this is the Murray ridge. The western arm, which starts near Socotra, is much shorter. From the point where the arms meet, the feature continues south-eastwards as the Carlsberg ridge. It consists of two ridges at depths generally of 2,300–3,000 m., with a minimum of 836 m., separated by a trough which reaches a depth of 3,383 m. South of Rodriguez the ridge is continued by the New Amsterdam and St Paul ridges towards the Kerguelen plateau and from there by the Antarctide ridge to Mt Gauss (90° E.).

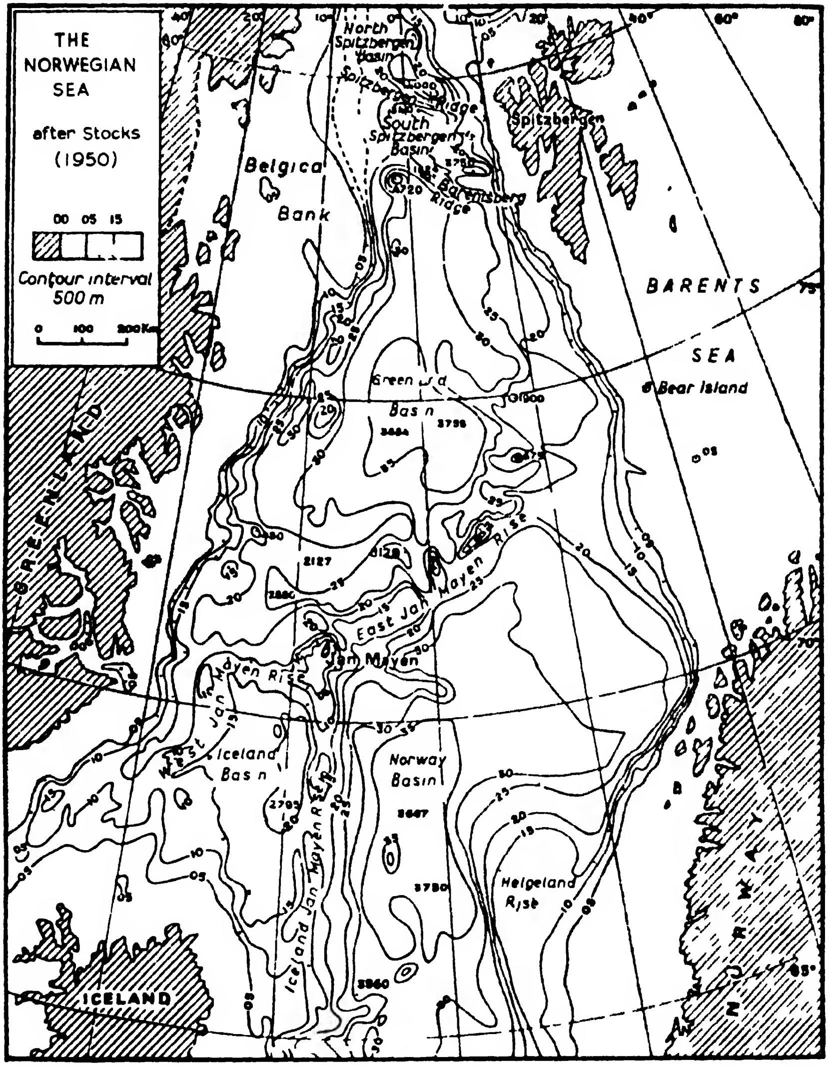


FIG. 35 THE NORWEGIAN SEA (AFTER STOCKS, 1950).

In both oceans comparatively minor ridges or rises separating the principal basins link up with the main ridge. The west Jan Mayen rise separates the Greenland and Iceland basins and is less than 1,000 m. deep in two places. Other examples are the Faroe–Iceland rise, which continues as the Wyville-Thomson ridge between the Faroes and Scotland, the Walfish Bay ridge, and the Cape ridge. These Atlantic ridges are usually north-east–south-west like the relief on the central ridge, but the Faroe Iceland rise lies north-west–south-east as do the two ridges joining Spitzbergen to the north of Greenland. In the Indian Ocean, the minor ridges form a curved herringbone pattern between the Carlsberg ridge and Madagascar. The Seychelles and Mauritius banks lie on the main ridge and Providence Island, Farquar Island, and the Cosmodelo group on another ridge.

In the Pacific there are many north-west–south-east ridges especially in the south under the Tuamotu, Society, and Tubuai Islands. Such ridges also occur in the north, where another set running west-south-west–east-north-east crosses them between Hawaii and the Marshall Islands. The flanks of the Hawaii ridge have plateaux down to depths below 1,200 m. In the south-east of the Gulf of Alaska, there are many small north-east–south-west ridges with depths of less than 1,000 m. in places. These ridges end southwards at the curious Mendocino escarpment, which extends from Cape Mendocino in California westwards as a straight line for at least 1,200 miles. The scarp faces south and is 3,000 m. high in places with slopes of 7–24° (Fig. 37 c). Farther south are three other large zones of highly irregular topography, running roughly parallel to the Mendocino escarpment: the Murray, Clarion, and Clipperton escarpments (Menard, 1955).

(c) *Seamounts and tablemounts.* Isolated seamounts, some of them several thousand metres high, are found in various places, as for example between Bermuda and the mid-Atlantic ridge, and between the Murray and Clarion escarpments in the Pacific. They have the same general form as an uneroded volcanic island. The tablemounts, or guyots as Hess named them after a nineteenth-century American geographer, are curious features first described by Hess in the north-west Pacific between Hawaii and the Marianne Islands. They are truncated cones rising 2,700–3,600 m. from the sea-floor and reaching 900–1,800 m. below the surface. Their summits are almost flat and are about 15 km. in diameter: relief on them does not exceed 70 m. The slope is gentle from the summit to the sides of the guyot, at least in the area where they were first described, but then steepens abruptly to 20°. Sometimes

the slope is terraced. Eniwetok atoll has been built between two guyots without obscuring their form. In the Gulf of Alaska, Menard and Dietz have also found guyots, the flat summits lying at depths of 700–2,070 m. (Fig. 37 A). Seapeaks, without truncated summits, lie near them. Guyots are also known in the north-east Pacific off California, where the flat summits are at depths of 720–500 m., in the south-west Pacific, and possibly in the north-west of the Indian Ocean.

Structural interpretation. The structure of the ocean floor can be deduced from the velocity of seismic waves passing through it, as the velocity is greater in the sima than in the sial. Where the bottom is free from sediment, data can be obtained by dredging. Most of the areas without sediment are the ridges and sills, so that dredging is mainly of use in these localities (p. 252). Seismic work, comparable with that done on the continental shelf, has been done on the central Atlantic ridge.

Dredging, seismic sounding, and the nature of the islands on the central ridges of the Atlantic and Indian Oceans, show that these ridges are mainly basaltic, but the basalts of the Carlsberg ridge are different from those of the Deccan. On the Atlantic ridge the Americans have found limestone, probably of Tertiary age, and have shown that the zone of plateaux on the west side of the ridge carries some 300 m. of sedimentary rocks between the parallels of 30 and 35 N. Quartzites have also been found on the southern part of the ridge. In the Indian Ocean the Seychelles and Socotra ridges are partly formed of granite. The Pacific ridges, as far as we can tell, are basaltic beneath the coral formations which surmount them. It will be recalled that granite and sedimentary rocks are sial and that basalt is sima, although basalt can be deposited over the sial during volcanic eruptions.

The seismic data, as recorded by several geophysicists, may be summarized as follows. The velocity of Love waves is greater on the floor of the Pacific and Arctic basins than beneath the continents; velocities on the floors of the Atlantic and Indian Oceans are intermediate between those of the Pacific and those of the continents. In the western Atlantic Ocean the velocities are of the same order as those on the floor of the Pacific, but they are much lower in the eastern Atlantic. In the Indian Ocean velocities are lower in the western part than in the eastern part.

From these facts the following conclusions have been reached by some workers. The Pacific, the western Atlantic, and the Arctic basin, at least in its central part, have sima at no great depth beneath the ocean floor. The eastern Atlantic and the western part of the Indian Ocean are

floored with sial, e.g. the Seychelles granite, although this is probably thinner than on the continents. This agrees with the distribution of gravity anomalies (cf. Chapter VII, p. 220). The true Pacific lies east of the Andesite line, which follows the outer edge of the Japan, Bonin, and Marianas trenches and then passes to the south between the Gilbert, Ellice, and Samoa Islands on one side and the Solomon, Fiji, Tonga, Kermadec, Chatham, and Macquarie Islands on the other. West of this line the floor of the ocean is sial and the dominant magma of andesitic type, this being produced through contamination with sial. Thus we may distinguish between true oceans, such as the Pacific east of the Andesite line, the western Atlantic and the Arctic basin, and continental oceans such as the western Pacific, the eastern Atlantic, and the western Indian Ocean. Of uncertain type are the eastern Indian Ocean, which probably belongs to the true oceans, and the east-south-east Pacific, between Easter Island and the Galapagos Islands, which appears to be intermediate between the true and continental oceans according to the *Capricorn* expedition of 1952–3.

It must be remembered that this distinction between true and continental oceans rests on an interpretation with which everyone would not agree, so that a number of reservations should be made.

Teixeira denies that a distinction should be made between the eastern and western Atlantic, as seismic data obtained by the Lisbon observatory shows that the velocity of Love waves in the eastern Atlantic is at least as great as in the west and of the same order as in the Pacific. Thus the east and the west Atlantic should both be regarded as true oceans. Recent work by Ewing's team and Hill and Laughton also supports the idea that the eastern Atlantic is basaltic. West of the Andesite line, in the south-west Pacific between New Zealand, Australia, and the Solomon and Samoa Islands, Officer thinks that the crustal thickness is not larger than in the southern Pacific, so that this part of the Pacific may be a true ocean as well as the central part.

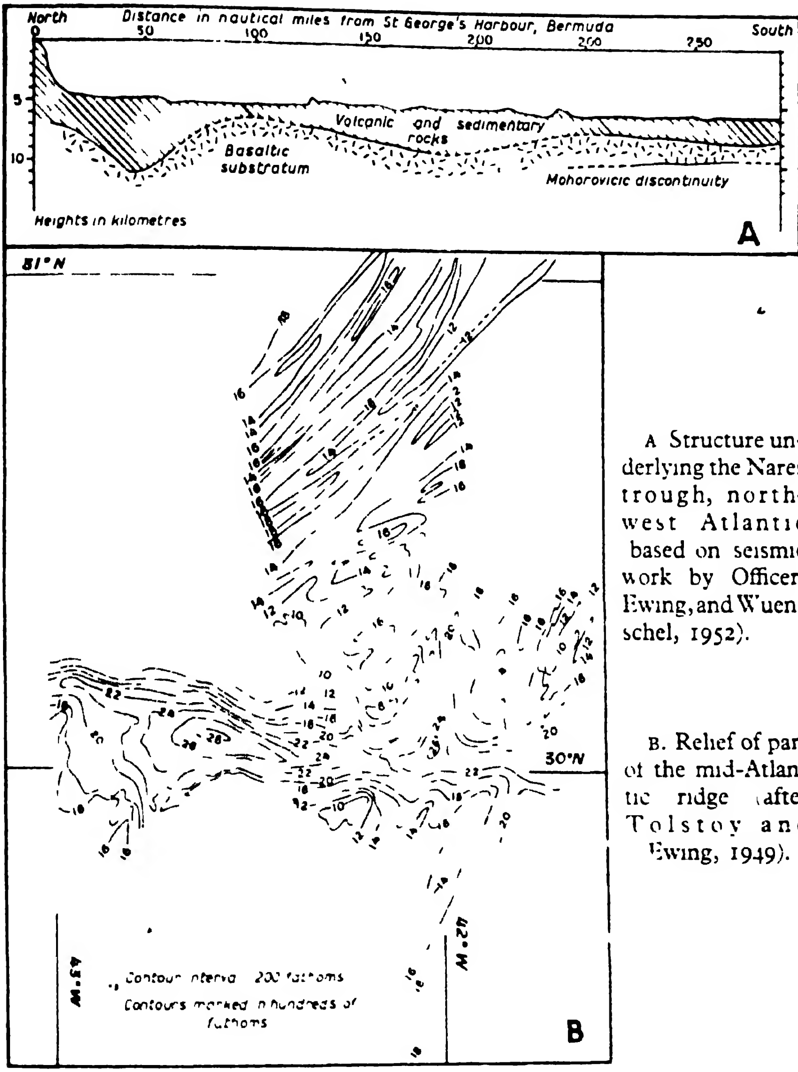
However, the distribution of relief in the eastern Atlantic agrees with its interpretation as a continental ocean. The ridges appear to be continuations of some of the major relief features of Africa: the Cape ridge appears to continue the Drakensberg escarpment, the Walfish ridge appears to be in line with the hills behind Mossamedes and the Congo–Zambesi watershed; the Guinea ridge is continued in a series of volcanic areas through Fernando Po, the Cameroons, Adamaua, and Hajjer el Khemis in Chad territory. In the Indian Ocean the Carlsberg ridge seems to be part of the Red Sea rift system which remains in part as a

trough and in part has been built up by outpourings of basalt: hence the difference between the trough of the Red Sea and the Carlsberg ridge projecting up from the ocean floor.

We must consider the justice of applying the term continental to the east Atlantic a little more fully. In places it has been the site of a sea for a very long time: a Senegal gulf has existed since the Upper Cretaceous, but in a broad view an epicontinental sea is negligible and does not radically affect the issue. Much more serious are the objections based on seismic soundings which, according to Weibull (quoted by Wiseman and Ovey), seem to indicate a great thickness of red clay in the deeper parts of the eastern Atlantic. As red clay accumulates so slowly the eastern Atlantic must have been in existence for a very long time, perhaps since the Silurian, and must have been very deep since red clay is characteristic of the deepest parts of the oceans. Nevertheless, the exact nature of the thick deposits on the sea-floor in this part of the Atlantic is still open to question.

We must also consider a little more fully the hypothesis that the floor of the Pacific is sima. East of the Andesite line, with the exception of the Galapagos region, there is almost general agreement that this has always been an ocean basin. But if this is so, a considerable thickness of sediment should have accumulated there. Kuenen estimated this at 3,000 m., which is still much less than the thickness of sial on the continents. This sedimentary layer beneath the Pacific may account for the subsidence of volcanic areas on which barrier reefs and atolls are built (see Chapter III), provided that the sediments have been slowly compacted beneath the weight of lava erupted on to them, and not as rapidly as the volcanic area is built up.

Similarly in the western Atlantic, there is certainly a covering, in places several kilometres thick, of sediment over the sima. In the Nares trough south-east of Bermuda, seismic work by Officer, Ewing, and Wuenschel has shown that up to 5,500 m. of sediments or volcanic rocks overlie the basaltic sima (Fig. 36 A). Nevertheless, the Mohorovicic discontinuity between the basalt and dunite layers is 9,000–12,000 m. below sea-level (3,000–6,000 m. below the ocean floor). This is much nearer to the surface than it is on the continents, where, beneath mountains, it may be 30 km. or more deep. In this way we may perhaps reconcile the Russian conclusions about the Lomonossov ridge in the Arctic, according to which this ridge is a part of a Mesozoic folded belt, and the American views (Oliver, Ewing, and Press) about the oceanic structure of the central Arctic.



A Structure underlying the Nares trough, north-west Atlantic based on seismic work by Officer, Ewing, and Wuenschel, 1952).

B. Relief of part of the mid-Atlantic ridge (after Tolstoy and Ewing, 1949).

FIG. 36 STRUCTURE AND RELIEF OF PARTS OF THE BOTTOM OF THE CENTRAL ATLANTIC

How do these ideas about the ocean floors fit in with Wegener's hypothesis of continental drift? The restriction of true ocean deeps to the western Atlantic is not in itself against the drift theory. It is possible that the separation of Europe and Africa from America may have taken place at the central ridge, because the well-known structural similarities between Africa and South America are closely paralleled by the similarities between the southern part of the ridge and the coasts of Brazil and Argentina. The difficulties are much greater on the Pacific coast of North America. According to Kuenen, any westward drift of the continent would have rucked up and folded the sediments on the floor of the Pacific, but the western cordilleras of America are not formed of deep-sea sediments.

It is difficult to reconcile the permanence of the Pacific with the enormous outpourings of lava which occurred there at the end of the Tertiary. The seismic stability of the area, which contrasts with the instability of its borders, does not accord with the opening up of enormous fissures on the ocean floor, whatever the mechanism responsible for these fissures, whether it be tear-faulting, overthrusting, or tension faulting resulting from folds.

It is not necessary to imagine an ancestral Pacific of greater depth than the other oceans. Although it contains more deep trenches and fewer ridges and sills than the other oceans, the depths in the basins are not much greater than those of the other oceans.

Finally, we must be very cautious in accepting the hypothesis of two basic types of oceans.

The parallel crests and troughs, characteristic of some ridges, may be folded ranges, especially where sedimentary rocks have been dredged up, e.g. the mid-Atlantic ridge. According to Dietz and Menard, the the plateaux on the flanks of the Hawaii ridge may be erosion forms submerged as a result of volcanic subsidence. The same authors think that the elongated crests in the Gulf of Alaska are late Tertiary or Quaternary folds by analogy with adjacent forms on the continents. But we are by no means sure of this.

The problem of seamounts and guyots must be considered separately. The seamounts may be submarine volcanoes which have never appeared above the ocean surface, or which subsided so quickly that they are virtually uneroded. In 1946 Hess put forward several explanations of guyots, but these were pure hypotheses, and it was only in 1951 and 1952 that some concrete data were provided by dredging. Dredging was carried out on two guyots off the Californian coast (Carsola and Dietz),

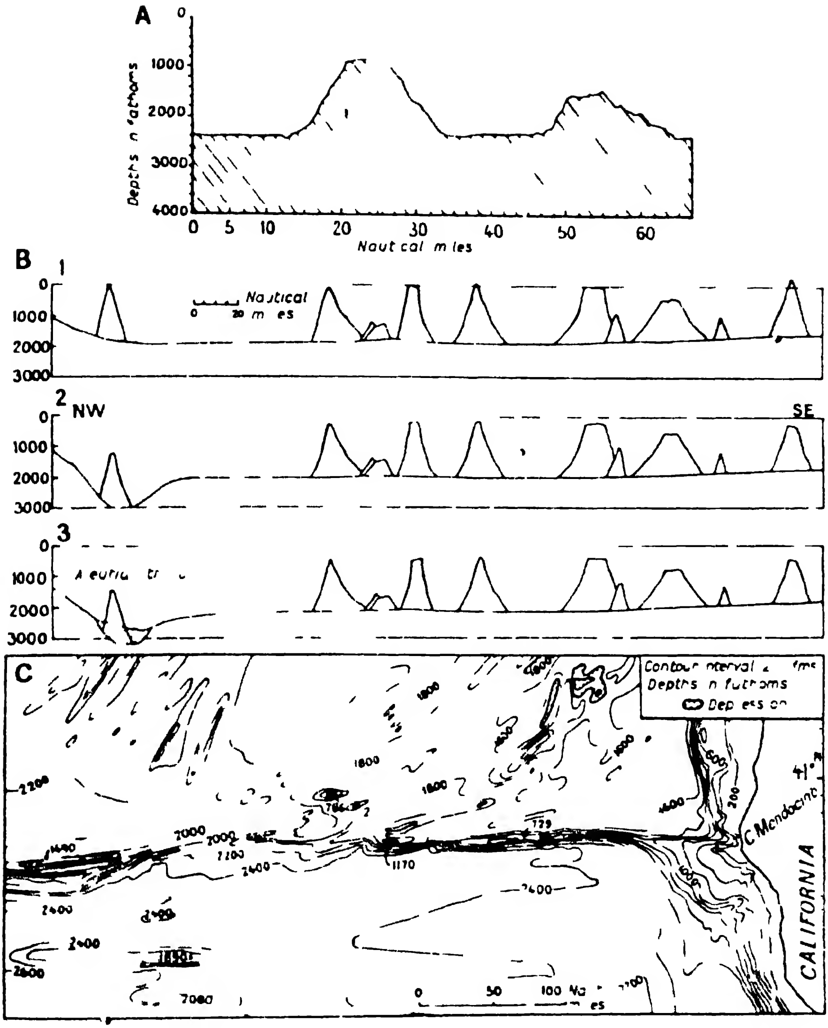


FIG. 37 RELIEF OF PARTS OF THE PACIFIC FLOOR

A Cross-section of a guyot (*left*) and a less well-defined seamount (*right*) in the Gulf of Alaska (after Menard and Dietz, 1951). B Presumed evolution of submarine mounts, peaks and guyots, in the Gulf of Alaska (after Menard and Dietz, 1951). C The eastern third of the Mendocino escarpment (after Menard and Dietz, 1952).

and on those between Hawaii and the Marianas Islands, which had previously been explored by Hess (Hamilton). On the flat summits of the former, which were actually photographed, bare rock is exposed at the surface except for several patches of Globigerina ooze in fissures. This forces us to invoke the action of underwater currents to keep the surface clear of sediment. In addition, boulders of basalt and basalt encrusted with manganese dioxide were found. The basalt is of the normal type found in the true Pacific. The boulders were so well rounded that they are most likely to be the remains of a beach. On the other group of guyots rounded basalt pebbles were again found together with basaltic sands and limestone resembling that forming today on the beaches of atolls. The latter was fossiliferous and included several types of coral, rudistids, gastropods, and one echinoid: it is a shallow-water fauna of Middle Cretaceous age. At the top of the flanks of the guyots were forms resembling coral banks.

These are the facts on which any hypothesis must be based. The guyots are most likely to be old eroded volcanoes, as Hess thought in 1946. This means that they must have reached the surface at some time. But we can no longer believe that they never supported coral reefs even in warm seas. Hess, who adopted this idea, was forced to regard them as pre-Cambrian, i.e. as being formed before any corals existed. The guyots in the western part of the central Pacific were probably volcanic cones truncated by the sea, partly covered with coral during the Cretaceous, and later subject to rapid subsidence which took them below the level of coral growth. On the other hand one of the guyots in the Gulf of Alaska is unusually deep (2,500 m.) and lies in the Aleutian trench, which leads us to suspect that this guyot was formed before the trench, i.e. before the Tertiary. The mechanism responsible for the rapid subsidence of the guyots is probably to be found in the overloading of the ocean floor caused by the volcanoes. The subsidence in tropical seas must have been too rapid for coral growth to keep pace with it, cf. the relative rates of subsidence and coral growth postulated by Darwin in his explanation of atolls. Fig. 37 B illustrates the proposed evolution of the guyots in the Gulf of Alaska.

B. TRENCHES AND ASSOCIATED FEATURES

Morphological characters

The trenches demand special attention although they occupy only about 1 per cent. of the total sea area.

They attain surprising depths: 10,863 m. in the Marianas trench, 10,790 m. in the Tonga trench, 10,504 m. in the Philippine trench, and 10,377 m. in the Kuril-Kamchatka trench. Each seems to consist of a series of closed basins, and they are all narrow, long, and generally curved. Their distribution is peculiar as they are always near the margin of the oceans, the majority being on the Asiatic side of the Pacific between the Aleutians and the Kermadec Islands. The six trenches not in this zone include three off the coasts of Peru and Chile, which are not quite as deep as the others, the Java trench in the Indian Ocean, and the Porto Rico and South Sandwich trenches in the Atlantic. Except for those off Peru and Chile, they are associated with volcanic island arcs, which lie between them and the continental masses. Between these volcanic arcs and the continents are basins up to 2,600 m. or more in depth, e.g. the Caribbean and New Britain basins. There is often a broad uplift of the ocean floor between the trenches and the open oceans (Fig. 38 A). In some cases, the pattern is more complex. There may be two associated island arcs, only the one farthest from the trench being volcanic as in the Ryukyu archipelago. Indonesia is even more complex (Fig. 38 B). The arrangement here from the ocean to the continent consists of the ocean basin, an outer trench, a non-volcanic island arc, an inner trench, a volcanic island arc through Sumatra and Java, another basin, and the old shield of Borneo. The additional complications present in the east of Indonesia and the Philippines will be dealt with later.

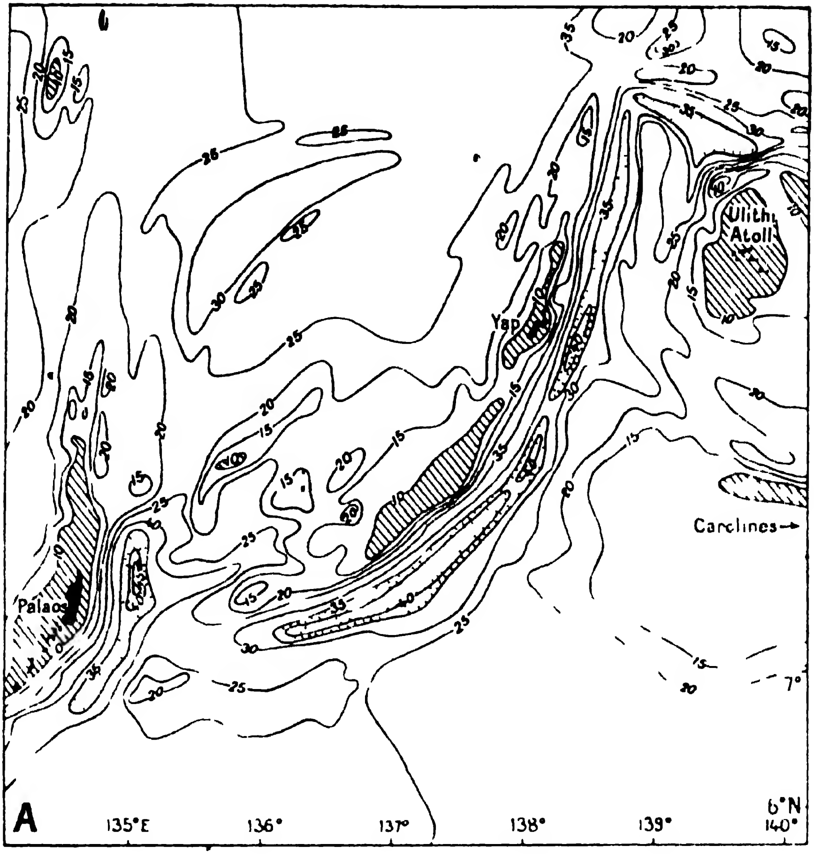
Geophysical evidence

Both gravity anomalies and the positions of the deep foci of earthquakes provide useful evidence.

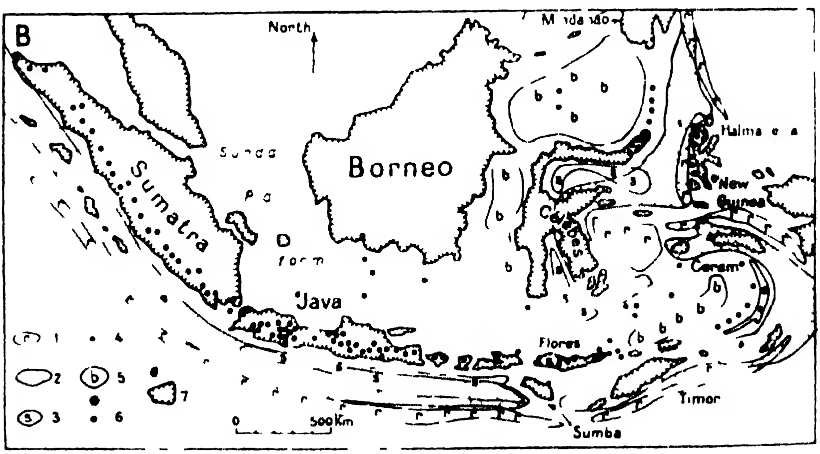
Gravity anomalies (Fig. 39 D) are sufficiently well know n a number of places for us to recognize two patterns corresponding to different morphological types.

The normal type shows a negative anomaly over the trench or on its inner flank and a positive anomaly on the island arc. This type occurs in the Antilles, Yap, Guam, and from the Bonin Islands to Hokkaido. It corresponds with the sequence—simple trench, volcanic arc, inner basin.

The complex type of Indonesia includes a positive anomaly over the ocean and the outer part of the trench, a negative anomaly on the non-volcanic arc, and a positive anomaly on the inner deep and the volcanic arc.



Ulithi Atoll
Yap
Palaos
Carolines →
A
135°E
136°
137°
138°
139°
140°
6°N
7°



B
North
Borneo
Sumatra
Java
Celebes
Halmahera
New Guinea
Ceram
Flores
Timor
Sumba
0
500 Km

Furthermore, all deep-seated earthquakes (300–700 km.) are associated with the trenches, with the single exception of the Hindu Kush ones. These earthquakes, and the shallower ones associated with them, do not have their foci on a vertical plane but a surface inclined at 30–55° from the horizontal, sloping down from the bottoms of the trenches towards the continents. This has been observed all round the Pacific including the Chile-Peru area. For example, the earthquakes associated with the Kurile arc originate under the Sea of Okhotsk between the Soya Strait and the west coast of Kamchatka. The zone of instability has the form of an oblique cut which descends from the bottom of the trench in simple examples to a depth of 700 km., i.e. one-tenth of the radius of the earth (Fig. 39 D). The depth of the trenches is thus a mere fraction of the depth of the discontinuities at the end of which they occur.

Interpretation

There is no point in summarizing all the older theories about trenches, since they were formulated before adequate geophysical data had been obtained. In the present state of knowledge, the best explanation of trenches is perhaps provided by the theories of Umbgrove and Vening Meinesz.

In the simple type, the negative anomaly on the trench shows that there is an abnormal thickness of low-density sial there. This agrees with the general picture of true and continental oceans given above, because the Andesite line, which marks the limit of the true Pacific, passes to the east of the trenches. This concentration of sial is thought to be related to a sinking of the crust where the oblique shear-zone meets the ocean floor (Fig. 39 A): the shear-zone may be due to the overriding of the continents, to some sort of flow structure or to convection currents in the interior of the earth. This downwarping of the sial is the basic cause of geosynclines or tectogenes. The sima is forced away from the trench, especially inward towards the continent. The sima forced

FIG. 38 THE TWO MAIN TYPES OF DEEP TRENCHES

A. Simple type near Yap and Palaos (after Hess). Contour interval, 500 fathoms. *From right to left*: raised part of ocean floor in front of trench; trench; volcanic arc; inner basin.

B. Complex Indonesian type (after Umbgrove and Vening Meinesz). 1. Outer trenches. 2. Negative gravity anomaly on outer non-volcanic arc. 3. "Intramontane" troughs. 4. Volcanoes mainly on inner arc. 5. Inner basins. 6. Earthquake foci below 500 km. 7. Land.

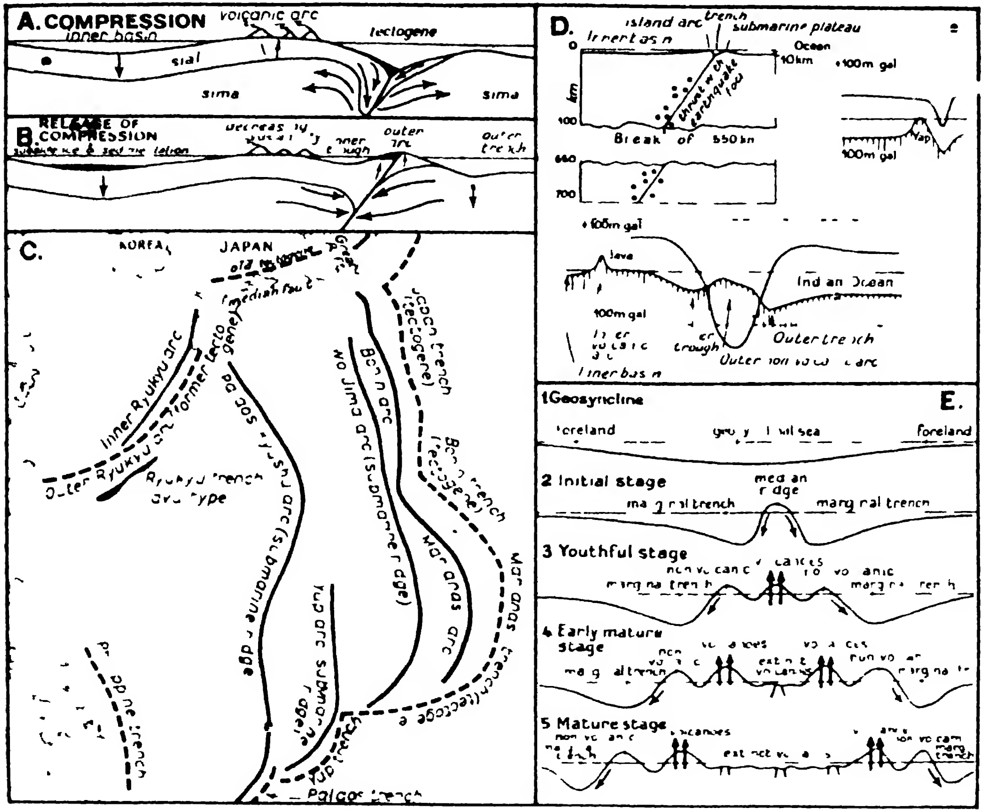


FIG. 39 ORIGIN AND ARRANGEMENT OF OCEAN TRENCHES

A and B. Main stages in the formation of trenches and island arcs (after Umbgrove). C. General arrangement of arcs and trenches off the Philippines (after Hess, except for Ryukyu). D. *Top left*: oblique thrust beneath a trench (after Benioff). *Top right*: gravity anomalies over simple type. *Bottom*: gravity anomalies over complex type (after Vening Meinesz). E. Formation of arcs and trenches according to van Bemmelen's theory.

towards the ocean results only in the slight upheaval of the ocean floor outside the trench. The bulk of sima which is forced landwards domes up the sial on the ocean floor there until it appears above the surface of the sea. This causes both the positive gravity anomaly and the appearance of the island arc, ruptures in which are followed by outpourings of andesitic lavas. Inside the arc there is a relative subsidence in the inner basin, as the sima is spread out and less active. As the arc rises, subaerial erosion acts on it and submarine landslips may occur. This leads to a thickening of the sial on the inner side of the trench, which is responsible for the fact that the negative anomaly is sometimes on this side of the trench and not in its centre. This theory also explains why coral reefs west of the Andesite line are often uplifted, while those to the east have subsided for reasons stated above. Simple patterns of this sort are found in the Japan, Bonin, Marianas, Yap, and Tonga-Kermadec trenches, and possibly in the Philippine trench. We need more gravity data before we can classify the New Britain, New Hebrides, Kurile, and South Sandwich trenches.

The Chile and Peru trenches are somewhat anomalous. They are concave towards the oceans, whereas all the others are convex except the Tonga Kermadec trench, which appears to be straight. They are relatively shallow and there is no associated island arc. But the concavity may be explained by their being nipped between two continental blocks, as the Pacific is probably not a true ocean in this part. The Andes take the place of the volcanic island arc. As there is a deep-seated thrust zone, they may be included in the simple type.

We can arrive at a satisfactory explanation of Indonesia, if we assume that it has passed the simple stage and has reached a more complex phase of evolution (Fig. 38 B and 39 B). Here the outer non-volcanic arc corresponds with the outer trench in the simple type as is shown by the negative gravity anomaly, for the outer trench of the complex type has a positive anomaly. The outer non-volcanic arc represents a mountain range rising from the geosyncline. Thus, if we neglect complications such as the folding of sediments in the inner basin, we may assume that a geosyncline was originally formed by compression and that when this pressure was released the sima moved back and forced up the geosynclinical sediments of the tectogene as a non-volcanic arc. The release of pressure is also accompanied by a phase of quiescence of the volcanoes of the inner arc. The external trench is, therefore, not a true trench from the geophysical point of view, as it is not situated at the end of an oblique thrust zone.

Gravity anomalies are also very useful in the interpretation of the complex area of eastern Indonesia between Celebes and New Guinea. Measurements made by Vening Meinesz, and the position of deep-seated earthquake foci, allow a comparatively simple interpretation of the area. The puzzling features are (1) the way in which the Philippines link up with the rest and (2) the volcanoes of Halmahera which are on the wrong side of the tectogene. The sinuous form of the latter may be due to compression between the continental masses of Sunda and Melanesia.

The probability of the theory is increased by the presence in Timor of deep-water sediments: Fig. 38 B shows that the island is in the zone of negative anomalies and that the rocks were originally in the bottom of the trench.

Similar considerations seem to lead us to class the Lesser Antilles as intermediate between the simple and Indonesian types. The negative anomalies are on the trench in front of the volcanic islands, and on Barbados, which contains deep-water sediments. Barbados, like Timor, would thus represent a range rising from the geosyncline. But here Ewing and Worzel do not agree with Vening Meinesz's interpretation: according to their measurements, the sial is not thick beneath the Caribbean Sea, the Mohorovicic discontinuity lying at only 12,000 m. under it and at 16,000 m. under the trench. For the Indonesian archipelago, the validity of Vening Meinesz's explanation has not been questioned by these American scientists.

For the Aleutian area, another possible explanation, including two oblique shear-zones on both sides of the island ridge, was recently put forward by Gates and Gibson, who gave a very accurate description of the submarine topography of this region.

We must finally examine the complex which stretches from Japan to the Palaos and to the Philippines (Fig. 39 C). This area is unique because there are two parallel series of trenches and volcanic areas. The outer stretches through Japan, Bonin, the Marianas, Yap, and Palaos, the inner through the Philippines and Ryukyu. Further, between the island arcs of the outer trenches in the north and the inner trenches, is a series of ridges which either break the surface or are at no great depth below it. Towards the south these become the island arcs associated with the outer trenches, and thus overlap the arcs associated with those trenches in the north. We have seen that all the outer trench island-arc systems are simple, while the Ryukyu would be of Indonesian type. We may, therefore, regard the outer systems, which develop successively

and not along one line, as more recent than the inner systems. There may be a third older and completely filled tectogene through Korea and Shantung. It is, however, unlikely, as one could expect, that the Philippines are of Indonesian type.

The theories stated above are not the only ones which attempt to explain island-arc systems. We must draw attention to the theory of van Bemmelen, which depends to some extent on the earlier ideas of Stille, Kober, and Kraus. It has recourse not to thrusting but to geochemical processes associated with volcanicity, which cause vertical and not horizontal movements. According to this theory mountain building and volcanicity start in the centre of the geosyncline, which later becomes quiescent when two lines of mountain building and volcanicity break out on the flanks and move outwards (Fig. 39 E). This theory has the advantage of giving a better explanation of certain features in the Philippines, as Luzon includes an internal volcanic arc and a non-volcanic arc on its Asian side running through Palawan. But the theory presents two great difficulties. It does not explain why trenches occur only outside the arcs, since it regards the structure as symmetrical, unless the inner basins are taken as the equivalent of the outer trenches, which is unsatisfactory as the form and depth of the two are both very different. It also leaves unexplained the inclined thrust surfaces on which the earthquakes originate and this is an essential part of these structures. On the whole the theory of Umbgrove and Vening Meinesz seems much more satisfactory.

The trenches, which almost take our study into the earth's interior, may also lead us back to the continents, for the negative anomalies of the Indonesian geosyncline are continued through the Nicobar and Andaman Islands into Burma; and the Aleutian trench ends in Yakutat Bay in Alaska. Finally, the presence of island-arc mountains on the edges of the trenches, which make a difference in elevation of 12,600 m. in 125 km. off Mindanao, shows the essential continuity of the land and ocean relief.

C. PELAGIC DEPOSITS

Sedimentation in relation to environment

Beyond the bathyal zone on the continental slope lies the abyssal zone, which includes the deep parts of the oceans and the trenches. But depth is not the only nor the main factor affecting sedimentation. The distance from the land is also of great importance. We may separate an hemipelagic-abyssal environment, lying within 200 or 300 km. of

the land, from a pelagic-abyssal environment, farther out to sea. In the former zone the effect of the land is shown more strongly in the nature and calibre of the sediments, so that there is a much greater chance of meeting terrigenous sediments even as coarse as sand. In the deeps of Indonesia, which all lie near the land, terrigenous sediment is abundant. Thus we may find coarser sediments in the deep trenches than in much shallower areas of the Pacific far away from land. It is, therefore, dangerous, though sometimes possible, to deduce the depth at which sediment was laid down from its facies: hence, we have not used the title 'abyssal sedimentation', for this section. In general, distance from the land is more important than depth, although the two often go together.

The pelagic environment is not entirely one of still water. Banks far from land are affected by waves: hence, depth is also important. But even in bathyal or abyssal conditions, there may be appreciable movement in the sea. Guyots and ridges, as Kuenen says, are frequently covered with coarse material such as sand, and even gravel, at depths of several hundred metres. In Indonesia the bottoms of some straits are swept by strong currents at depths of 1,000–2,000 m. so that only solid rock and shingle are left there. Slower but more widespread currents flow on the sea-bottom and explain why the deposition of fine material does not take place even on little-marked sills. Off south California, according to Shepard, a corer was broken on hard rock at a depth of 3,600 m. Rock occurs on the bottom near the Canaries at 2,000 m.

Submarine landslides, mudflows, and turbidity currents from the continental slope may also affect the deep areas and give rise to mixed and abnormally bedded sediments. Deposits, thought to have been derived from the Hudson canyon by Ericson, Ewing, and Heezen (p. 217), have been found at depths of 4,370–4,940 m.

According to the same authors and Kuenen, the 1929 earthquakes on the Grand Banks of Newfoundland triggered off a great slide on the continental slope, which became a turbidity current and travelled more than 450 miles, breaking submarine cables on the way. Some doubt was raised by Kullenberg and Shepard about the reality of such a current off the Grand Banks, and, in the latter's opinion, landslides provide a better explanation for this and for other submarine features described by him in front of the Mississippi delta. But the turbidity current hypothesis has been put forward again more recently to explain the break of five cables on the floor of the Mediterranean, following the violent earthquake near Orléansville, Algeria, on 9th September

1954. Turbidity currents are a good explanation of sandy layers reported by Bourcart from a deep-sea core here. Moreover, soundings have shown that on the ocean floor south-east of Ceylon and also off California there are long shallow channels, sometimes in line with submarine canyons, bounded by ridges comparable with the levées of normal rivers. American authors tend to regard these as the result of turbidity currents, some of the material in suspension having been deposited at the sides.

Icebergs also deposit erratic material of all sizes when they melt, and the sea-floor round Antarctica is dotted with morainic material. In core samples from these seas the glacial phases may be recognized from the material brought by the icebergs. Data from the north of the Ross Sea have shown that the glaciations correspond with those of the northern hemisphere. Large pieces of volcanic pumice may float a considerable distance before sinking and thus be found far from the land.

But the exceptions must not be over-emphasized. The pelagic zone below a few hundred metres suffers little movement: it is essentially the zone of fine sedimentation as a result of distance from the land (for settling velocities, see p. 24): coarse material occurs here only as the result of peculiar conditions, and will sooner or later be covered with fine sediment.

An unusual environment is the euxinic or Black Sea type, where stagnation leads to a lack of oxygen and the presence of only anaerobic bacteria at depth. The Black Sea itself is the best example, but certain Indonesian basins such as the Sulu basin and the south of the Celebes basin possess somewhat similar conditions. All areas of this type are not pelagic, since some occur in the depressions in the Norwegian fjords very near the land.

Types of deposits

There are scarcely any terrigenous deposits, apart from red clay, which are peculiar to the areas considered here. The bottom often has a high proportion of such material, but it is not so very different from other deposits which occur nearer the coasts at shallower depths.

For instance, blue mud, which becomes grey on drying, is met at all depths from 0 to somewhat less than 5,000 m., especially in the Atlantic where it covers 4 million sq. miles, in the Pacific (3 million sq. miles), and the Southern Ocean (2½ million sq. miles). Terrigenous material on the average makes up 75 per cent. of the whole, but tends to decrease with depth.

Black mud, the colour being due to iron sulphide and organic matter, also has a large vertical distribution. At depth, it is associated with the sort of conditions experienced in the Black Sea, and may be considered as an extreme form of coastal and estuarine mud. In the Black Sea sapropelitic muds contain 30 per cent. of unreduced organic matter. These muds contain much hydrogen sulphide and consequently stink. Hydrocarbons are formed in such areas.

Volcanic muds are common round Indonesia, where there are many active volcanoes. The *Snellius* expedition showed that they predominate over large areas in the Celebes basin and in the Flores depression to below 5,000 m. and even beyond 300 km. from the land. They are found in smaller quantities much farther from volcanoes: it is well known that the 1883 eruption of Krakatoa spread ash over the whole of the earth's surface. Submarine lava flows, which cool and consolidate in the form of pillow-lava, may be the cause of certain rock-floored deeps.

More curious are the deep-sea sands, which may contain up to more than 50 per cent. of fine or even medium sand (1 mm.) formed mainly of terrigenous elements. They are found mainly in the Atlantic and down to below 7,000 m. in the deep Romanche trench. Unworn sand found at depths of about 4,300 m. near the Canaries is interpreted by Mlle Duplaix and A. Cailleux as the result of a series of volcanic eruptions. Sands off the coast of Sahara, which often show as great a degree of wind abrasion as the sand of the desert of Mauretania (Duplaix and Cailleux), could be explained by wind action but for the size of the grains, which range up to 0·7 mm. They may be due to turbidity currents or submarine landslides carrying to great depths material originally deposited in shallow water by sandstorms. The sands in the Romanche trench, which are very angular, have been explained by certain authors as the result of the fragmentation of abyssal rocks by tectonic forces (?).

But these terrigenous deposits only cover a small part of the ocean floors. Organic deposits and red clay are much more widespread and cover 74 per cent. of the whole ocean floors. They include calcareous Globigerina and Pteropod oozes and siliceous Radiolarian and Diatom oozes (Fig. 40). The calcareous or siliceous nature depends on the nature of the skeletons of the organisms forming the deposits.

Globigerina ooze covers 35 per cent. of the ocean floors and is made up of the skeletons of foraminifera. It is the most widespread deposit in the Atlantic, where it covers half the ocean floor, excluding the adjacent

seas but including the neritic and bathyal zones of the Atlantic proper. It is equally predominant in the south-east and south-west of the Pacific, and in the west of the Indian Ocean. Globigerina ooze is defined as consisting of more than 30 per cent. Globigerina tests: thus some Globigerina ooze may contain predominantly terrigenous material and some predominantly red clay. The calcareous content averages 65 per cent. and may reach 97 per cent. Part of the calcium carbonate seems to be a chemical precipitate. Globigerina ooze is rarely encountered below 5,000 m.

Pteropod ooze is rare and covers only 1 per cent. of the ocean floors. It covers a narrow north–south strip in the middle of the south Atlantic and several small areas off Brazil and in the north Atlantic. Pteropods rarely form 30 per cent. of the deposit by weight. The calcareous content averages 74 per cent. It is found between 1,500 and 3,000 m.

Radiolarian ooze is almost as rare and covers only 1–2 per cent. of the ocean floors, principally in an east–west strip north of the Equator in the Pacific. It contains 20–70 per cent. of Radiolarian tests. The calcareous content varies between 4 and 20 per cent. It is characteristic of great depths.

Diatom ooze, covering 9 per cent. of the ocean floors, is found in high southern latitudes and in the north of the Pacific, as it is a cold-sea deposit. It contains 3–25 per cent. of mineral matter, largely derived from floating ice. Diatoms sometimes make up more than 90 per cent. of the deposit. According to the *Valdivia* reports, the calcareous content is 2·7–24 per cent.

Red clay, which covers 28 per cent. of the ocean floors and must not be confused with the red mud of the continental shelf of the Guianas, is essentially a deep-sea pelagic deposit. It covers most of the north Pacific, the centre of the south Pacific, the deeper parts of the main Atlantic basins and depressions, and the east of the Indian Ocean. It usually, therefore, occurs in the deepest parts of the oceans. From the samples obtained there it may also occur in the Arctic basin, apart from the Lomonossov ridge where the deposits are coarser. It is finer than the organic oozes (it contains 85–86 per cent. of material less than 0·05 mm. in diameter, whereas Radiolarian ooze contains 40 per cent. Diatom ooze, 16–20 per cent., Globigerina ooze, 28–31 per cent., and Pteropod ooze, 20 per cent.). In addition it contains manganese nodules, which may exceed 10 cm. in diameter bones of whales, pumice, material dropped by glaciers and floating trees, and a few tests of foraminifera, Diatoms, and Radiolaria. The average calcareous

content is 7–10 per cent. and the maximum 29 per cent. Spectrographic analysis has shown that the bulk of the clay is not volcanic or meteoric matter, as had been previously thought, but clay minerals. These are

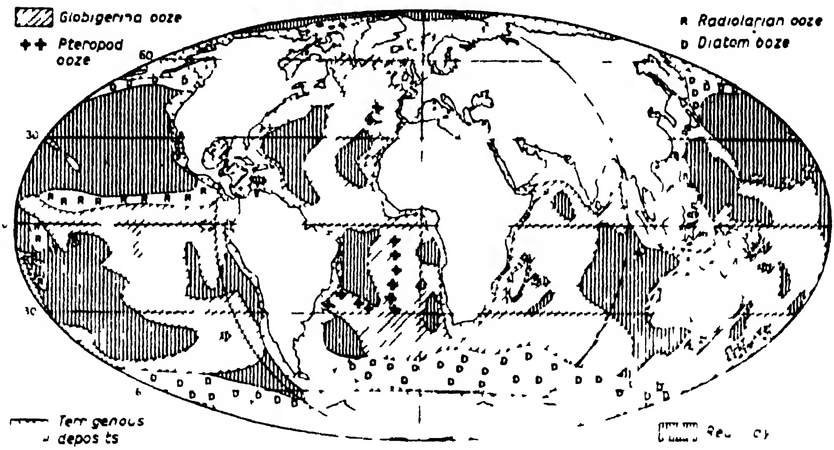


FIG. 40 DISTRIBUTION OF PELAGIC DEPOSITS

probably derived from subaerial weathering and did not settle nearer land because they were so fine. Red clay is thus a terrigenous deposit of a peculiar character.

The following is the distribution and average depth of organic oozes and red clay, according to Sverdrup, Johnson, and Fleming

	Indian Ocean (%)	*Pacific Ocean* (%)	*Atlantic Ocean* (%)
Calcareous ooze	54·3	36·2	67·5
Siliceous ooze	20·4	14·7	6·7
Red clay .	25·3	49·1	25·8
Total .	100·0	100·0	100·0

	Calcareous ooze (%)	*Siliceous ooze* (%)	*Red clay* (%)
Indian Ocean	26·9	33·9	15·7
Pacific Ocean	40·6	55·3	68·7
Atlantic Ocean	32·5	10·8	15·6
Total .	100·0	100·0	100·0

	Average depth (m.)
Pteropod ooze	2,072
Globigerina ooze	3,612
Diatom ooze	3,900
Radiolarian ooze	5,292
Red clay	5,407

Rate of deposition

The rate of deposition may be determined from cores, provided that the age of the material is known. The age may be determined in several ways: by the recognition of glacial phases in the cores; recognition of the ash of eruptions of known date (this method was used in the Moluccas); from the presence of annual varves (this has been done in the Black Sea and the Gulf of California); and radioactive measurements of the sediments, etc. The compression caused by the corer has also to be taken into account. Shepard and Kuenen estimate this at about 40 per cent., so that a 3-m. core represents about 4 m. of unconsolidated sediment on the sea-floor, but only 2·4 m. of consolidated rock into which it will be transformed with time.

The rate of sedimentation, especially of red clay, in the deepest parts is generally extremely slow. The various methods used give rates of roughly the same order. According to Kuenen the rate of deposition of unconsolidated red clay is 0·4–1·3 cm. per 1,000 years, depending on the methods of age determination; for Globigerina ooze it is 0·8–4 cm.; and 0·7 cm. for Diatom ooze, the last being based on the recognition of glacial phases. Shepard, from a variety of sources, gives 0·6 cm. per 1,000 years for red clay and 1 cm. for Globigerina ooze. The rates vary with locality even if the same method is used but the maximum rate of accumulation of Globigerina ooze in the rth Atlantic cannot exceed 4 cm. per 1,000 years. The rates only increase greatly when there is much terrigenous material, other than red clay. In four localities in the Flores depression the *Snellius* expedition found an average accumulation of 23 cm. in 115 years, since the eruption of Tampora in 1815, i.e. a deposition of 2 m. per 1,000 years; but at one of the four stations Globigerina ooze would have made up only 0·9 mm. of this 1,000-year total. This area is transitional to the neritic areas where sedimentation is even more rapid, e.g. 5 m. per 1,000 years at the entrance to the Clyde. The great ocean depths are obviously not, therefore, the sites of future mountain chains, material for which is accumulated either on subsiding continental shelves, or on the

continental slope, or at the edge of the deep trenches, where subaerial erosion, corals, submarine landslides, and finally subaerial volcanoes provide abundant material.

Problems of distribution

The distribution of organic oozes may be partly explained by the environment preferred by the organisms which form them. Globigerinae live mainly in warm water. Diatoms frequent cold water and withstand freezing in certain cases. Radiolaria require warmer water than Globigerina. Hence the presence of ooze derived from them in the zone of the North Equatorial Current of the Pacific Ocean.

Because of the temperatures to which different species of foraminifera are adapted, oozes derived from them are of great importance in the study of the Pleistocene. Most of these creatures live at depths of less than 100 m. Thus the climatic changes of the Pleistocene should have affected the water where they live and be reflected in the foraminiferal deposits of the sea-floors. Interpretations of this sort have been made of cores taken from low latitudes in the north and south Atlantic, the Caribbean, the Indian Ocean, the Arctic Ocean, etc. One core taken with a Kullenberg corer from the Caribbean at a depth of 4,700 m. showed eleven or twelve changes from tropical to temperate conditions in 15·40 m. These are attributed to Pleistocene glacial phases. The second interglacial is probably represented by a warm fauna extending over a considerable part of the middle of the core. Comparable results have been obtained from other seas. This is one of the most promising fields of oceanography, for, by it, it may be possible to establish which sills were uncovered by the sea and so determine the extent of the eustatic changes of sea-level. However, the effect of submarine landslips must be guarded against in the interpretation of cores.

But the distribution of the two most important organic deposits, Globigerina and Diatom oozes, is probably not only related to the relative abundance of these organisms in the sea. The way in which a deposit principally of red clay is formed is also a problem, because the rate of deposition of organic oozes is generally much greater.

We meet here a further aspect of the problem of the solution of calcium carbonate in sea-water, already discussed in the first chapter (pp. 27–9). To explain the distribution of Diatom and Globigerina oozes, Sverdrup, Johnson, and Fleming invoke the following processes. It has been shown that most of the bottom water of the Pacific and Indian Oceans and much of that of the Atlantic is derived from the

Antarctic, where it is formed at the surface, sinks and spreads as far as low latitudes. This water passes over the Diatom ooze first and dissolves the calcareous tests deposited there: thus the siliceous Diatom oozes are thought to be residual deposits. As the water moves to lower latitudes, it contains more and more calcium carbonate. By the time that it reaches the Globigerina areas it is saturated or even to some degree supersaturated with calcium carbonate, so that Globigerina ooze can accumulate in middle and low latitudes without the calcareous tests being dissolved.

The same authors offer a general explanation for the almost complete absence of calcareous tests in the red clay. In the Pacific, the deep circulation is very slow, from the south to the north and then back to the south, for there is, in contrast to the north Atlantic, practically no sinking of surface water in the north Pacific to spread out into low latitudes. During its long slow movement the oxygen content of the water decreases and its CO_2 content increases from south to north. Thus, beyond a certain point, calcareous tests can be dissolved as the result of the abundance of CO_2: hence the scarcity of Globigerina ooze and the abundance of red clay and Diatom ooze in the northern half of the Pacific. The smaller proportion of red clay in the Atlantic is thought to be caused by the higher lime content of Atlantic water: the reason for this remains to be discovered. If the contrast in lime content really exists, the distribution of Globigerina ooze and red clay in the Atlantic is partly explicable, for red clay is confined to the deepest parts, where solution, as Kuenen says, has the greatest chance of occurring for several reasons: the longer the time taken for a test to fall to the bottom, the greater is the chance of its being dissolved, while the low temperature, low salinity, and probably the increase in pressure will all help the action of solution. These may also be the reasons for the fact that siliceous oozes are found at greater average depths than calcareous oozes (see table on p. 257).

The problems of the chemistry of sea-water will probably be solved only gradually, because the methods of investigation are slow and difficult and many observations are required. In this, as well as in other aspects of oceanography, there is need for team work and for the comparison of results by specialists well versed in their techniques.

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